By implementing a series of mass-conserving nested high-resolution models down to approximately 1 km resolution that have realistic bathymetry, coastline, wind forcing and river run-off, the winter 1996–1997 shelf flow near Cape Mendocino, California, is simulated and compared with available observations from the Strata Formation on Margins (STRATAFORM) marine geology program. The model simulations are used to statistically characterize winter circulation, and to associate shelf flow response with specific forcing mechanisms. Major forcing mechanisms include highly complicated bathymetry in the vicinity of the headland of Cape Mendocino, and winds that are controlled by propagating storm systems. These storms have high wind speeds (up to 20 m s⁻¹) and drop significant precipitation that causes flooding of the rivers that drain the coastal mountain range.

Nesting techniques developed in meteorology are first tested in idealized scenarios that extract crucial dynamics of the coastal ocean. Realistic high-resolution nested models are then embedded in the 9-km resolution Naval Research Laboratory Pacific West Coast (NRL PWC) regional model of the North Pacific Ocean. The first nested model has 3 km resolution and extends from
38.6°N to 43°N, and offshore to 127.1°W. This model displays strong alongshore variability with more energetic flow located north of Cape Mendocino. This alongshore asymmetry is enhanced by river run-off that bathes the shelf north of Cape Mendocino during winter storms. An anticyclonic eddy north of Cape Mendocino is a generic circulation feature that occurs as poleward winds weaken and change direction during storm passage.

The 1-km resolution nested model, extending from 39.5°N to 42°N, and west to 125.9°W, is used to simulate the 80-year January 1997 Eel River flood event. The simulation illuminates the interaction of the Eel River plume with the background shelf flow. The plume responds to wind forcing in a manner consistent with Ekman dynamics. Furthermore, the plume becomes entwined in the mesoscale eddy whose formation and characteristics were documented in the 3-km simulation.
Modeling Studies of the Coastal Circulation off Northern California

by

Julie D. Pullen

A DISSERTATION

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in partial fulfillment of
the requirements for the
degree of

Doctor of Philosophy

Completed March 27, 2000
Commencement June 2001
I understand that my dissertation will become part of the permanent collection of Oregon State University libraries. My signature below authorizes release of my dissertation to any reader upon request.

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Julie D. Pullen, Author
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MODELING STUDIES OF THE COASTAL CIRCULATION OFF NORTHERN CALIFORNIA

1 INTRODUCTION

Recent in situ and satellite-derived observations of the California Current system have revealed rich mesoscale activity within that eastern boundary current (Huyer and Kosro, 1987; Huyer, et al., 1998; Strub and James, 2000). The seasonal development consists of an equatorward jet formed in the spring and summer upwelling season that migrates offshore and spawns mesoscale eddies in the arms of its meanders during fall. In winter, the mesoscale activity moves farther offshore and poleward currents dominate inshore (Lynn and Simpson, 1987). These observations characterize the larger-scale circulation that extends offshore many hundreds of kilometers.

Embedded in the California Current system, equatorward flow over the shelf during spring and summer is linked to equatorward (upwelling-favorable) winds. The Coastal Ocean Dynamics Experiment (CODE, 1981–1982) region from approximately 38°N to 39°N was chosen largely for the absence of significant forcing apart from the wind. Current meter mooring array design emphasized resolving the across-shore structure of shelf current response to wind forcing (Lentz, 1991). Analytical and observational work stemming from CODE have established the importance of local and remote wind forcing for driving coastal circulation during
the upwelling season (Chapman, 1987; Denbo and Allen, 1987). Other more recent observational programs were designed to examine the response of shelf flow to other forcing mechanisms.

Largier et al. (1993) found that open-ocean forcing impacted shelf circulation during the spring and summer Northern California Coastal Circulation Study (NCCCS) of 38°N to 42°N during 1987-1989. Furthermore, they found much alongshore variability in shelf response as well as evidence for small-scale motions in the wind band (periods of 1.6 to 19 days) that were unresolved by the 50-100 km alongshore mooring line separation. Barth, et al. (2000) describe observations of an equatorward upwelling jet interacting with the coastal promontory of Cape Blanco, Oregon during the Coastal Jet Separation (CJS) experiment. Hickey, et al. (1991) document a current created by buoyancy forcing from the Fraser River that flows against the prevailing (upwelling-favorable) equatorward winds off Vancouver Island.

More recently, an Office of Naval Research-sponsored observational program to study sediment deposition, Strata Formation on Margins (STRATAFORM), has instrumented the northern California coastal area in the vicinity of 40.6°N (Nittrouer, 1999). A goal of the STRATAFORM program is to understand the dynamics of sediment transport once the sediment leaves the Eel River and enters the coastal ocean. As sediment delivery is most intense during the winter storm season, instrument deployment is centered around that time. Northern California winter coastal currents are influenced by a variety of forcing mechanisms: strong and highly variable winds, buoyancy and momentum input from the Eel River, and alongshore variability in the coastline and bottom topography.

Current meter observations have been used to characterize seasonal aspects of shelf circulation. Due in part to the steadiness of upwelling circulation, obser-
vational analysis focused first on describing the well-surveyed upwelling regime and then on contrasting it with the less-observed winter flow. Huyer et al. (1978) described summer shelf flow off Oregon as possessing a baroclinic mean, while current fluctuations were barotropic in nature. By contrast, winter flow had a barotropic mean with baroclinic fluctuations. Lentz and Chapman (1989) extended this comparison to northern California using observations from CODE. They found the mean flow to be barotropic (baroclinic) in winter (summer), but found fluctuating currents were baroclinic independent of season. This pattern also applied to the NCCCS observations off northern California (Largier, et al., 1993). The initial categorization of winter flow in terms of its relation to the well documented spring/summer flow has recently shifted to analyses that focus on the complexity of wintertime shelf current structure. Dever (1997) found CODE winter across-shore velocities to be responsive to episodic wind events. Lentz and Trowbridge (1999) discovered a common vertical time-mean current structure during fall and winter at several mid-shelf sites sampled during various field programs off northern California (including CODE and NCCCS). Along-shore flow was poleward in the mean at all depths with maximum velocities of 5–10 cm s\(^{-1}\) in the middle part of the water column.

Though there is a growing catalog of in situ observations allowing common flow properties at different sites to be synthesized (Lentz and Trowbridge, 1999), in situ observations are necessarily local in scope and not usually temporally coherent. Furthermore, important circulation takes place at scales not resolved by the spacing of current meters. Remote sensing, though broad in spatial scope and temporally coherent, only provides information on surface ocean characteristics. Though large-scale geostrophic currents can be inferred from satellite altimetry, these observations become less reliable within 30 km of the coast (Strub and
Therefore there exists a place for numerical models, carefully verified against existing observations, to elucidate spatial and temporal characteristics of flow that cannot be resolved by observations.

Oey (1996), Masson and Cummins (1999), and Gan and Allen (2000) conducted high-resolution (less than 5 km) numerical studies of local coastal circulation on the West Coast. Reflecting observational emphasis on the upwelling season, these modeling studies explore the dynamics associated with specific flow features during that season. Oey (1996) studied the mechanism of eddy formation in the Santa Barbara Channel in summer using the Princeton Ocean Model (POM) with 5/3 km resolution. Oey (1996) concluded that the advection of vorticity produced at the corner of the coastal bend generated the eddy. Masson and Cummins (1999) used a realistic configuration of POM with 3 km resolution. They examined the depth-integrated dynamical balances leading to a poleward buoyancy-driven coastal current off Vancouver Island. Gan and Allen (2000), using a 1-km resolution POM simulation, investigated the relaxation response from upwelling-favorable wind forcing near Point Arena, in the CODE domain. They present momentum term balances to clarify the role of enhanced pressure gradient forces in the coastal flow reversal associated with relaxation. These studies demonstrate the importance of local high-resolution models in illuminating dynamical features of shelf flow response especially when complicated forcing like wind reversals, coastal promontories and buoyancy input are operative.

Local high-resolution modeling studies are within the range of present computational resources. It is desirable to achieve high resolution in a local area while still retaining information about the surrounding ocean. The technique of nesting was developed in response to this need. In nesting, information from a coarse resolution model is supplied continuously at the boundary of a nested,
high-resolution model. This allows for larger-scale features to influence the evolution of the flow field on the nested high-resolution grid without those features being simulated directly by the high-resolution model.

The field of meteorology has preceded ocean modeling in developing and implementing techniques to conduct communication between grids of differing resolution. Some of the techniques developed in those fields include conservative interpolation to guarantee mass conservation between the two models (Clark and Farley, 1984) and radiation nesting to provide for the removal of small-scale noise that might be produced at the nested grid boundary (Carpenter, 1982). These sophisticated techniques have primarily been developed and tested in mesoscale meteorological models designed for weather forecasting (Grell et al., 1995; Clark et. al., 1996). Nesting in primitive equation ocean models has not to date incorporated these advanced techniques. These techniques will be described thoroughly in a subsequent chapter. However we can look to prior nested modeling studies in the fields of climate and ocean modeling to provide guidance about the choice of nested domain size as well as nested domain resolution.

Climate modeling studies beginning about ten years ago have used coarse resolution General Circulation Models (GCMs) to supply boundary conditions to fine resolution Limited Area Models (LAMs) in order to facilitate regional climate prediction. Jones, et al. (1995) found that the synoptic scale of the nested grid was unconstrained by the coarse grid when the nested grid domain was large. By contrast, the small domain nested grids were fairly well-constrained. These results indicate that a choice of nested grid domain should be influenced by the spatial scales that are the intended subject of the simulation. Jones, et al. (1995) chose a domain size that was fairly constrained but yet showed little boundary distortion. By comparing surface air temperature and precipitation patterns, Jones, et al.
(1995) establish that on scales resolved by the GCM, the nested LAM matched
the GCM and observations, and that additionally, the nested LAM captured the
fine-scale detail found in the observations.

In ocean modeling, Fox and Maskell (1996) recently conducted a simulation
of frontal gradients, meanders and eddies in the region between Iceland and
Britain. They used a 9-km coarse grid and 3-km nested grid. Based on idealized
simulations, Fox and Maskell (1996) anticipated difficulties when the nested grid
boundary intersected steep topography or fronts. In order to test the robustness
of the boundary treatment the authors initialized the nested model with fine-
scale observational data and compared it to the original run initialized with the
field interpolated from the coarse grid. In contrast to their idealized simulations
(Fox and Maskell, 1995), initializing with features that the coarse grid could not
resolve did not lead to distortions or spurious activity originating at the boundary.
In realistic simulations of a month duration the nested model created a tighter
Iceland-Faroe front and more meandering and eddy activity compared with that
displayed by the coarse resolution model. Pulsed overflow through a gap in the
topographic ridge due to the meandering of the Iceland-Færøe front was found in
observations as well as in the nested model. This time-varying overflow of cold
water did not appear in the coarse grid simulation.

Oey (1998) used nesting to investigate the model resolution required to re-
solve eddies and meanders in the Iceland-Færøe frontal region. By increasing the
nesting ratio from 3 to 5 to 7 inside a 10-km resolution coarse grid, Oey (1998)
showed that resolution of the front required at least 10/3 km resolution; while
frontal meanders possessing scales on the order of 10 km required less than 2 km
resolution.
Nesting seems to be a viable means of gaining high resolution in a local area without neglecting the dynamics in the encompassing ocean. However, the grid resolution required to properly resolve features of interest depends on scales intrinsic to the phenomena under scrutiny. No prior nesting studies have been conducted for upwelling/downwelling processes on the continental shelf. Two-dimensional subkilometer resolution studies of upwelling and downwelling shelf circulation have found that solution convergence is attained at resolutions below 1 km (Allen, et al., 1995; Allen, et al., 1996). For three-dimensional shelf flow this limit has not been tested.

The object of this thesis is to study shelf circulation at high resolution using a series of nested models. Chapter 2 discusses the challenges and opportunities afforded by nesting. Next several nesting methods borrowed from meteorology are introduced. The nesting methods are tested in simple flow scenarios representative of governing continental shelf processes. Nesting is conducted between model domains of 3 km and 1 km resolution.

The implementation of a series of realistic nested models in a regional model of the North Pacific Ocean is detailed in Chapter 3. The region of enhanced resolution is from 38.6°N to 43°N, containing the NCCCS and STRATAFORM observational domains. The model includes the major forcing mechanisms: wind, buoyancy and bathymetry/coastline. Chapter 3 then describes a realistic 100-day 3-km resolution simulation of winter 1996–1997 shelf flow on the nested grid. The choice of the winter season makes the work unique among high-resolution modeling studies of the West Coast. The study characterizes the shelf response to a series of passing storms in terms of statistical quantities. The model is compared with observations from the NCCCS and STRATAFORM experiments.
An anticyclonic eddy north of Cape Mendocino is identified as a generic response pattern to intense winter storms.

Chapter 4 focuses on a 1-km resolution study of the 80-year January 1997 Eel River flood. The domain of this simulation lies within the domain of the 3-km resolution model and extends from 39.5°N to 42°N. The Eel River plume response to variable wind-forcing is described and compared with observations from STRATAFORM. The plume interaction with the shelf flow is documented as the severe storm progresses.
2.1 ABSTRACT

Several sophisticated nesting techniques borrowed from the field of numerical weather prediction were applied to an idealized primitive equation model configuration of the continental shelf. The study was designed to test the ability of the nesting approaches to accurately reproduce two-dimensional and three-dimensional upwelling dynamics without creating distortions at the nested grid boundary. A model of 1 km resolution was embedded in a periodic channel with 3 km resolution. The evolution of flow features on the nested grid was compared quantitatively and qualitatively against solutions from a benchmark 1-km resolution grid over the whole domain. A “radiation nesting” technique designed to permit boundary reflections to propagate out of the nested domain performed better than a method where boundary values were fixed or “clamped” at values interpolated from the coarse grid. A higher order diffusion operator improved the quality of the solution using the clamped technique.

2.2 INTRODUCTION

Coastal circulation encompasses a wide range of spatial scales. Energetic mixing processes over the shelf are significant at subkilometer scales in the surface and bottom boundary layers. Upwelling/downwelling fronts on the continental shelf are characterized by widths of several kilometers (Allen et al., 1995; Allen et al., 1996). The alongshore coastal jet that develops as a response to upwelling and the poleward undercurrent both extend at least the Rossby radius (typically 10 to 30 km) across-shore (Kosro et al., 1991; Pierce et al., 2000). Furthermore, recent observational evidence suggests that mesoscale eddies and jet meanders impinge on the continental shelf and slope (EG&G, 1991). Eddies and meanders
in the California Current that influence the coastal region have typical scales of 30–100 km.

Computational constraints pose a significant challenge to the simulation of high-resolution domains. Nesting is a finite-difference modeling technique to simulate a high-resolution grid of a limited area domain embedded in a lower-resolution grid of a larger scale domain. Because of the broad range of scales that characterize the coastal environment, nesting would seem to be an ideal means of modeling small-scale dynamics for a limited area while allowing the large scales generated on the coarse grid to influence the nested grid.

In nesting it is necessary to transfer values of variables from a coarsely spaced grid to a finely spaced grid at the location of a boundary or boundary region (Fig. 2.1). The boundary of the nested grid where the values are prescribed is an "artificial boundary" — that is, it does not correspond to any physical boundary. We can expect that errors of varying severity will be generated at the nested grid boundary due to overspecification (giving more boundary data than the equations require) and by supplying values that may be dynamically incompatible with the nested grid values (Perkey and Kreitzberg, 1976). The goal of nesting is to minimize severe distortions at nested grid boundaries while generating dynamics in the interior of the nested grid that are similar to the dynamics that would be obtained on a uniformly fine grid.

Used by scientists intent on numerical weather prediction since the 1970's, nesting has only recently been implemented in the ocean realm (Oey and Chen, 1992; Fox and Maskell, 1996; Oey, 1998). These simulations, though groundbreaking, did not take advantage of some of the more sophisticated nesting techniques that have been developed in meteorology. Such techniques include methods to conserve mass between the nested and coarse grids and radiation
FIGURE 2.1 Schematic of the nested grid configuration. On this C-grid, $u$ (across-shore velocity), $v$ (alongshore velocity), and $t$ (representing the scalars potential density, surface elevation, and the turbulence quantities) are locations where the nested grid receives interpolated values from the coarse grid.
conditions to render more permeable the transition region between the coarse and nested grid.

Clark and Farley (1984) imposed a conservation constraint on the interpolation from the coarse to the nested grid. Specifically, the conservation condition:

$$\sum_{\epsilon=-N^2}^{N^2} \Phi(\epsilon) \Delta l = \Phi_0 \Delta L$$  \hspace{1cm} (2.1)

is preserved, where $\Phi$ (or $\varphi$) are arbitrary coarse (nested) variables, $\Delta l$ is the horizontal nested grid spacing, $\Delta L$ is the coarse grid spacing, and $N = \Delta L / \Delta l$ is the nesting ratio. A variational problem is solved to derive an interpolation procedure that is consistent with the conservation condition (See Appendix). As shown in Fig. 2.2 for the case of $N = 3$, conservative interpolation satisfies (2.1) while linear interpolation does not. Using conservative interpolation guarantees that if a field is interpolated and then averaged, the original field is obtained exactly. When velocity transports are interpolated in this manner, mass fluxes in the continuity equation are consistent between the two grids.

Typically, nested grid values are updated at the boundary location with values interpolated from the coarse grid. Instead of being viewed as a barrier fixed at a value prescribed by the coarse grid, the nested grid boundary can alternately be viewed as a permeable membrane. When framed as an open boundary problem, nesting can benefit from the application of open boundary radiation schemes that have been developed for limited area models (Røed and Cooper, 1986). A hybrid method of nesting that incorporates a radiation condition is termed "radiation nesting" here. First suggested by Carpenter (1982), the nested grid boundary is updated using a radiation condition that "radiates away" the difference between the coarse variable and the nested variable. Heuristically, unwanted energy due
FIGURE 2.2 Comparison of interpolation methods in one dimension for a nesting ratio of 3. The numbers given in the legend for the interpolation methods are $C_0 = \frac{\Phi(-1/3) + \Phi(0) + \Phi(1/3)}{3}$. For the interpolation technique to be conservative, the quantity $C_0$ must equal the value of the underlying coarse variable, $\Phi_0$. Interpolated values at other locations are also indicated. Fine grid values interpolated by the linear method are connected with a solid line while those interpolated with the conservative method are linked with a dashed line. Note that the solid line (linear) overlays the coarse grid values exactly while the dashed line (conservative) does not. The linear interpolation, however, does not obey the conservation condition ($C_0 = \Phi_0$) expressed more generally in (2.1).
to the mismatch at the boundary between coarse and nested values is allowed to pass out of the nested grid domain. Mathematically,

\[
\frac{\partial \varphi_{\text{nest}}}{\partial t} = \frac{\partial \varphi_{\text{coarse}}}{\partial t} - c \frac{\partial (\varphi_{\text{nest}} - \varphi_{\text{coarse}})}{\partial y}
\]  

(2.2)

where \( \varphi \) is a generic variable and subscript \( \text{coarse} \) denotes a variable interpolated from the coarse grid. The computational phase speed \( c \) is the speed of the unwanted energy \( (\varphi_{\text{nest}} - \varphi_{\text{coarse}}) \) and can be computed using the method described by Orlanski (1976) (See Appendix). Theoretically, spurious reflections do not pile up, but are allowed to pass through the permeable boundary established by the radiation condition. Chen (1991) demonstrated how the radiation nesting procedure can eliminate acoustic and gravity wave reflection in an atmospheric model. Kunz and Moussiopoulos (1995) used this technique in atmospheric wind simulations. Perkins et al. (1997) introduced this nesting approach to the ocean modeling community, but did not implement it between grids of differing resolutions. Perkins and Smedstad (1998) implemented a form of radiation nesting for a limited area open ocean domain within a layered model of the Mediterranean Sea.

Though there is a substantial body of literature documenting nesting implementations in numerical weather prediction (Koch and McQueen, 1987), the transfer of those methods to different sets of equations (having different physics as well as different numerical discretization schemes) and different simulation scenarios (having different time and space scales) is problematic. Thus, nesting schemes must be tested in flow scenarios that approximate the anticipated application in order to ensure that major features of the flow environment are not compromised by the nesting strategy.

No previous nesting study has focused on circulation processes on the continental shelf with a high-resolution (on the order of 1 km) inner grid. As the first
step in implementing nesting in a regional model of the North Pacific Ocean, the
ability of various nesting procedures to capture important physical features of the
coastal domain is tested. These features include wind-forced Ekman transport
in the surface layer, bottom boundary layers over sloping bathymetry, and up-
welling fronts that may intersect the nested grid boundary. The nesting methods
evaluated include conservative interpolation and a radiation nesting condition.
In evaluating these methods, the utility of nesting is quantified in simple ide-
alized model configurations that reproduce essential aspects of continental shelf
circulation.

2.3 MODEL CONFIGURATION

The stratified, hydrostatic primitive equation Princeton Ocean Model (POM)
of Blumberg and Mellor (1987) with linear slope bathymetry is employed in this
modeling study. The coarse grid model has 3 km horizontal resolution, while the
nested grid model has 1 km resolution. The coarse model domain is a periodic
channel of dimensions 90 km by 90 km. The nested domain extends approxi-
mately 40 km offshore and 55 km alongshore. Both models use 30 vertical sigma
levels. No normal flow and free-slip are used as sidewall boundary conditions.
The maximum ocean depth is 400 m.

Simulations are carried out with a constant Coriolis parameter of 9.5 x
$10^{-5}$ s$^{-1}$. In these simulations, potential density ($\sigma _\theta$) replaces temperature and
salinity. The initial stratification profile on both the coarse and nested grids is
linear with $N^2 = 1.5 \times 10^{-4}$ s$^{-2}$. Time steps are 5 seconds for the external mode
and 150 seconds for the internal mode. Time steps on the coarse and nested grid
FIGURE 2.3 Model domain and wind stress for a.) alongshore uniform simulation, and b.) alongshore variable simulation.
are equal. Upwelling-favorable wind forcing is applied to the initially linearly stratified coastal ocean.

In the first set of experiments a spatially uniform wind is ramped over 1 day to 0.5 dyne cm\(^{-2}\) and held at that value for the duration of the 10 day simulation (Fig. 2.3a). In the second set of experiments the wind stress is uniform in the across-shore direction and has a sinusoidal variation in the alongshore direction. The wind stress is ramped over three days to a maximum magnitude of 1.0 dyne cm\(^{-2}\), then turned off for 2 days (Fig. 2.3b).

Two nesting techniques are tested here. The first fixes the boundary of the nest with values interpolated from the coarse grid. Termed "clamped nesting" here, this one-way nesting is accomplished by supplying values interpolated from the coarse grid to the nested grid boundary. Insertion of linearly interpolated coarse grid values at the nested grid boundary has been employed in nesting applications in ocean modeling (Oey and Chen, 1992; Fox and Maskell, 1996; Oey, 1998). Here, clamped nesting is carried out using linear interpolation on all prognostic variables, and alternately using the conservative interpolation technique of Clark and Farley (1984) on all prognostic variables (with velocities interpolated in transport form). The second technique uses the radiation nesting condition of Carpenter (1982) (See Appendix). In the radiation boundary condition, flow variables from the coarse grid are interpolated using the conservative technique of Clark and Farley (1984) with velocities interpolated in transport form.
2.4 ALONGSHORE UNIFORM SIMULATION

In this set of experiments the wind and bathymetry are both uniform in the alongshore direction. Therefore the model results should be two-dimensional with only across-shore and depth variations being permitted.

However, by inserting the nested grid we introduce the possibility that mismatches between the solutions on the 3-km and 1-km grids will contaminate the nested grid solution. Skamarock and Klemp (1993) warn that grid-scale-dependent parametrizations (e.g., Smagorinsky diffusion and turbulence closure schemes) that function at one scale may not perform in the same manner when the resolution is changed. When applied over a wide range of scales, grid-scale-dependent parametrizations may imperil solution convergence. This could cause boundary specification at the nested grid interface to be even more problematic. Since requiring two-dimensionality in the nested solutions is a stringent criterion, we simplify solution compatibility issues by employing the Richardson number-dependent turbulence formulation for vertical diffusivity and kinematic viscosity originally developed by Pacanowski and Philander (1981). Allen et al. (1995) compared the differences in the continental shelf flow response for the Pacanowski-Philander scheme and the Mellor and Yamada (1982) level 2.5 turbulence closure scheme. They found reasonably similar dynamics generated by the two schemes. To remove grid-scale-dependence in horizontal diffusivity and kinematic viscosity we set the coefficients of the second order operator constant and equal at $10 \text{ m}^2 \text{s}^{-1}$ on both the coarse and nested grids.

In coastal modeling, a desirable horizontal grid spacing for accurately simulating shelf processes is 1 km. With the aim of attaining 1 km resolution on a nested grid embedded in a regional model, the idealized studies discussed here use a 3-km resolution model as the "outer" model. Previous two-dimensional
studies of continental shelf circulation by Allen et al. (1995) used less than 1 km resolution. In investigating solution convergence here, two-dimensional solutions of 1 km, 1/2 km, and 1/6 km resolution displayed very good agreement amongst themselves. However, the 3-km solution diverges from the 1-km and higher resolution solutions.

Flow variables at day 10 on the fine (1 km resolution) grid of the alongshore uniform simulation are shown in Fig. 2.4 (left column). The upwelling-favorable winds drive offshore flow in the surface Ekman layer. On day 10 the compensating onshore flow is primarily in the bottom Ekman layer (middle panel). The resultant warping of isopycnals (top panel) is accompanied by an alongshore jet in agreement with the thermal wind balance (bottom panel). The dynamics of the model two-dimensional upwelling solution is discussed in Allen et al. (1995).

The coarse grid solution (Fig. 2.4, right column), though dynamically similar, has important discrepancies compared to the 1-km solution. In the bottom boundary layer potential density ($\sigma_\theta$) has less structure on the 3-km grid compared to the 1-km grid. The frontal structure in across-shore velocity ($u$) in the surface Ekman layer is not well resolved on the 3-km grid compared to the 1-km grid. The alongshore jet has a smoother shape in the upper 30 m on the 3-km grid compared to the 1-km grid. These differences can be sources of error in the nesting procedure. In what follows, we will see to what extent the discrepancy between the solutions confounds the achievement of a purely alongshore uniform solution on the nested grid.

Near-surface nested grid solutions at day 10 using the clamped technique are shown in the left column of Fig. 2.5. Departure from alongshore uniformity is evident, particularly in across-shore velocity ($u$) where two grid point noise
FIGURE 2.4 Cross-sections of fine (1 km) and coarse (3 km) grid solutions on day 10 showing potential density ($\sigma_\theta$, in kg m$^{-3}$), across-shore velocity ($u$, in m s$^{-1}$), and alongshore velocity ($v$, in m s$^{-1}$).
reflected from the boundary severely contaminates the interior solution. (Using the mass-conserving interpolation does not improve the quality of the solutions.)

Some atmospheric nested models utilize higher-order diffusion operators (Grell et al., 1995) because a more scale selective operator helps in damping small-scale noise at the boundary, especially two grid point noise. We change the second order operator used for horizontal diffusion and kinematic viscosity to a fourth order operator on both the coarse and nested grids. The coefficient for the biharmonic operator on the coarse grid is $10(2\Delta X)^2$, where $\Delta X$ is the grid spacing. On the nested grid the coefficient is $10(2\Delta x)^2$ where $\Delta x$ is the grid spacing. The change to the more scale-selective operator (middle column, Fig. 2.5) does improve the quality of the nested solutions. However the solutions still differ (especially across-shore velocity) from two-dimensionality. (Similar results were obtained by increasing the (second order) horizontal diffusion coefficient from 10 m$^2$ s$^{-1}$ to 100 m$^2$ s$^{-1}$. However, the upwelling response was significantly weakened by this increased diffusion and the solution still did not conform to strict two-dimensionality.)

Solutions from the simulation using the radiation nesting condition described in the Appendix are presented in the right column of Fig. 2.5. The radiation condition represents a significant improvement in the quality of the nested grid solutions. The fields come very close to maintaining alongshore uniformity.

2.5 ALONGSHORE VARIABLE SIMULATION

In this scenario, the wind has a spatially varying sinusoidal shape in the alongshore direction (Fig. 2.3b). Furthermore, the wind is shut off after day 3. This simulation is intended to test how well nesting methods are able to handle
FIGURE 2.5 Comparison of alongshore uniform nested grid solutions on day 10 on sigma level 3. Shown are potential density ($\sigma_\theta$, in kg m$^{-3}$), across-shore velocity ($u$, in m s$^{-1}$), and alongshore velocity ($v$, in m s$^{-1}$).
advection of three-dimensional time-dependent features. Clamped nesting using conservative interpolation on velocity transports and all other prognostic variables is compared to radiation nesting. Horizontal diffusion and kinematic viscosity coefficients for the second order operator are $10 \, \text{m}^2 \, \text{s}^{-1}$ on both the coarse and nested grids. In order to approach more closely the simulations we plan to carry out in realistic coastal settings, this simulation utilizes the Mellor-Yamada (1982) level 2.5 turbulence closure scheme. Hence $q^2$ (twice the turbulent kinetic energy) and $q^2l$ ($q^2$ multiplied by the turbulence length scale) become prognostic variables that need to be interpolated from the coarse grid and communicated to the nested grid in the nesting procedure.

In order to assess the success of the nesting methods tested here, comparisons of the nested grid solutions are made against solutions from a 1-km grid over the whole domain. Both visual inspection of selected flow variables as well as statistical comparisons are shown. First the model response on the fine (1 km) and coarse (3 km) grids are compared; then the nested grid solutions are presented.

In the fine (1 km) resolution simulation a tongue of dense water is upwelled next to the coast in the surface layer during the wind forcing regime (Fig. 2.6, left column). Once the wind is shut off at day 3, this tongue is advected downstream by the decaying alongshore jet. The coarse grid (Fig. 2.6, right column) solution shows choppiness in the contour lines. In addition, at day 4 there is a closed contour that is represented on the fine grid but does not appear on the coarse grid. The nested grid potential density using the clamped method has spurious reflections at the southern boundary (Fig. 2.7, left column). The radiation nesting method (Fig. 2.7, right column) shows no such noise at the boundary. Both methods capture the closed contour at day 4 that was not reproduced by the coarse grid.
FIGURE 2.6 Potential density ($\sigma_\theta$, in kg m$^{-3}$) on sigma level 3 for the fine (1 km, left column) and coarse (3 km, right column) grids following the cessation of the wind.
FIGURE 2.7 Potential density ($\sigma_\theta$, in kg m$^{-3}$) on sigma level 3 on the nested grid after the wind has been shut off. (The nested grid region has been isolated and enlarged compared to Fig. 2.6.) Shown are solutions using clamped (left column) and radiation (right column) nesting.
FIGURE 2.8 As in Fig. 2.6 but for across-shore velocity \((u, \text{ in m s}^{-1})\).
FIGURE 2.9 As in Fig. 2.7 but for across-shore velocity ($u$, in m s$^{-1}$).
FIGURE 2.10 As in Fig. 2.6 but for along-shore velocity \( (v, \text{ in m s}^{-1}) \).
FIGURE 2.11 As in Fig. 2.7 but for along-shore velocity ($v$, in m s$^{-1}$).
The character of the oscillations that the across-shore velocity \((u)\) undergoes following the termination of the winds is similar on both the coarse and fine grids (Fig. 2.8). However, there are differences in shape and strength. There are severe reflections on the nested grid southern boundary using the clamped method (Fig. 2.9, left column). Such reflections are minimized on the nested grid that uses the radiation nesting (Fig. 2.9, right column).

On the fine and coarse grids alongshore velocity \((v)\) weakens in time after the wind has been shut off (Fig. 2.10). The nested grid using the clamped technique has unwanted noise at the southern (outflow) boundary (Fig. 2.11, left column). The transition across the boundary is smooth using the radiation nesting (Fig. 2.11, right column). Clamped nesting with boundary values interpolated linearly from the coarse grid performed at about the same level as the clamping using conservative interpolation presented here.

The rms error (relative to the fine grid) of nested across-shore velocity \((u)\) at three depths is shown in Fig. 2.12. The time-varying nature of the flow is evident in the pictures: the errors tend to change behavior and oscillate after day 3 when the winds are shut off. The radiation nesting had the lowest rms error when compared to the clamped and the coarse solutions. Other variables had less pronounced reduction in rms error using the radiation nesting.

Though the results described above utilized the same time step on both the coarse and nested grids, it is computationally advantageous to use a larger time step on the coarse grid. The nesting simulations described above were repeated with a time step three times larger on the coarse grid (both internal and external) compared to the nested grid. Boundary conditions for the nested grid were linearly interpolated in time from the coarse grid before they were incorporated in the nested model using either the clamped or radiation nesting technique. No
FIGURE 2.12 Rms error relative to the fine grid of across-shore velocity (u) at sigma levels 3 (top), 15 (middle), and 27 (bottom).
material differences were found from the results described above where the same time step was used on both grids.

2.6 DISCUSSION AND SUMMARY

We have conducted numerous tests of numerical formulations of radiation nesting (2.2) and we found (2.15) and (2.16) (see Appendix) to give the best results. Durran et al. (1993) have reported that the Orlanski-type approach to calculating computational phase speed is susceptible to boundary induced errors leading to dramatic nonphysical oscillations in phase speed. The modified Orlanski method of Carmelengo and O'Brien (1980) that is employed here sets an upper limit on the speed at which outward radiation can progress. This limits the magnitude of oscillations that can occur. Still, the work presented here would benefit from any advancements in the formulation of robust radiation conditions.

Though it is tempting to conclude that radiation nesting is a superior approach compared to clamped nesting, caution must be exercised in extending these results to realistic flow scenarios. Indeed, the idealized experiments of Fox and Maskell (1995) demonstrated that the presence of steep bathymetry or strong currents at the nested grid boundary generated grid-scale noise. However Fox and Maskell (1996) did not find such noise when applying their nested model to the steep bathymetry and strong velocities associated with the Iceland-Færøe front.

In addition, Kunz and Moussiopoulos (1997) found that for realistic wind simulations over Athens the radiation nesting approach required a much smaller time step than a non-radiative approach. In our preliminary nested simulations of the northern California coastal circulation (having high current speeds and very steep bathymetry) we have found that the radiation nesting technique requires
a prohibitively small time step. Even for the idealized simulations more variables must be saved to enact the inter-grid communication when using radiation nesting. (This includes values of prognostic variables at several spatial locations adjacent to the boundary as well as at several time levels.) So clamped nesting may yet prove to be the most practical approach for a broad range of realistic oceanographic conditions, especially when the nested grid can be configured so that the region of dynamical interest is located at a distance from the nested grid boundary.

Nesting was implemented in several simple flow scenarios that display features characteristic of continental shelf flow. In a simulation whose forcing had no variation alongshore, performance was assessed by requiring strict two-dimensionality of the solution. The one-way nesting techniques of clamping and radiation nesting were compared. Clamping caused distortion of the nested solution, which was especially severe for across-shore velocity. Use of a higher-order diffusion operator served to damp two grid point noise at the nested boundary but did not ameliorate deviations from two-dimensionality in the interior. Radiation nesting proved the most effective at approximating a two-dimensional solution on the nested grid.

In a simulation with alongshore variations in wind forcing, the radiation nesting condition gave superior results when compared to clamped nesting using a high-resolution simulation as a benchmark. Clamped nesting generated spurious reflections at the downstream boundary that were visible in all prognostic variables. However, these reflections did not significantly impact the integrity of the solution away from the nested grid boundary.
2.7 ACKNOWLEDGMENTS

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2.8 CONSERVATIVE INTERPOLATION

Starting from the conservation statement (2.1), Clark et al. (1996) construct a quadratic interpolation formula that is consistent with this averaging operator. A consistent formulation is achieved by deriving \( \bar{\varphi}(\varepsilon) \), a least square fit to \( \varphi(\varepsilon) \), subject to the constraint that (2.1) is maintained. The problem is posed using variational calculus:

\[
\Lambda = \sum_{\varepsilon=-\frac{1}{2}}^{\frac{1}{2}} \left[ \bar{\varphi}(\varepsilon) - \varphi(\varepsilon) \right]^2 \Delta t + \lambda \left[ \sum_{\varepsilon=-\frac{1}{2}}^{\frac{1}{2}} \bar{\varphi}(\varepsilon) \Delta t - \Phi_0 \Delta L \right],
\]

where

\[
\bar{\varphi}(\varepsilon) = a_0 + a_1 \varepsilon + a_2 \varepsilon^2,
\]

and \( \lambda \) is the Lagrange multiplier. In Fig. 2.2, \( \varepsilon = -1 \) corresponds to the location of \( \Phi_{-1} \), \( \varepsilon = 0 \) corresponds to the location of \( \Phi_0 \), and \( \varepsilon = +1 \) corresponds to the location of \( \Phi_{+1} \). In order that consistency is achieved over all nested grid cells lying within the coarse grid cell, the sum extends over the entire width of the coarse grid cell that contains the variable \( \Phi_0 \). Therefore \( \varepsilon \) spans the range \( \pm \frac{1}{2} \).

The standard procedure for solving the Lagrange multiplier problem (eq. 2.3) is to minimize \( \Lambda \) with respect to \( \lambda, a_0, a_1, \) and \( a_2 \):
and then solve for the quantities $a_0$, $a_1$, and $a_2$.

After some algebra, the interpolation can be expressed in a general form:

$$\bar{\varphi}(\epsilon) = E_- \Phi_{-1} + E_0 \Phi_0 + E_+ \Phi_{+1},$$

with

$$E_- = \frac{\epsilon(\epsilon - 1)}{2} + \alpha,$$
$$E_0 = (1 - \epsilon^2) - 2\alpha,$$
$$E_+ = \frac{\epsilon(\epsilon + 1)}{2} + \alpha,$$

where

$$\alpha = \frac{(\Delta l)^2}{24} - 1.$$

In implementing $\bar{\varphi}(\epsilon)$, the value of $\epsilon$ at the desired location of the interpolated variable is used. That is, in Fig. 2.2 (for $N = \Delta L/\Delta l = 3$), to get the value interpolated to the nested grid at the location of $\Phi_0$, $\epsilon = 0$ in (2.10) to (2.12). To obtain interpolated values for the two other nested grid cells lying within the coarse grid cell, $\epsilon = -\frac{1}{3}$ and $\epsilon = \frac{1}{3}$, respectively.

2.9 IMPLEMENTATION OF RADIATION NESTING

Consider the northern boundary of the model domain where $j$ is the location of the boundary and $j - 1$ is interior to the boundary. Rearranging the radiation nesting condition (2.2):
\[
\frac{\partial (\varphi_{\text{nest}} - \varphi_{\text{coarse}})}{\partial t} + c \frac{\partial (\varphi_{\text{nest}} - \varphi_{\text{coarse}})}{\partial y} = 0. \tag{2.14}
\]

Let \( \dot{\varphi} = \varphi_{\text{nest}} - \varphi_{\text{coarse}} \). Then the implicit finite difference leapfrog discretization of the radiation condition at the space location \( j - 1 \) and time level \( n + 1 \) is:

\[
\frac{(\varphi_{j-1}^{n+1} - \varphi_{j-1}^{n-1})}{2\Delta t} = -c_0 \left[ \frac{1}{2} (\varphi_{j-1}^{n+1} + \varphi_{j-1}^{n-1}) - \varphi_{j-2}^n \right]. \tag{2.15}
\]

Solving for \( c_0 \) gives:

\[
c_0 = \frac{\Delta y}{\Delta t} \frac{(\varphi_{j-1}^{n+1} - \varphi_{j-1}^{n-1})}{(\varphi_{j-1}^{n+1} - \varphi_{j-1}^{n-1}) - 2\varphi_{j-2}^n}. \tag{2.16}
\]

Discretizing (2.14) at location \( j \) and rearranging gives:

\[
\varphi_{(\text{nest})j}^{n+1} = \varphi_{(\text{coarse})j}^{n+1} + \frac{\varphi_{j}^{n-1} - \frac{c\Delta t}{\Delta y} (\varphi_{j}^{n-1} - 2\varphi_{j-1}^{n})}{1 + \frac{c\Delta t}{\Delta y}}, \tag{2.17}
\]

where \( c \) is determined following the "modified Orlanski" formulation of Carmelengo and O'Brien (1980). The condition was found to work better if it was used in the form:

\[
c = \begin{cases} 
\frac{\Delta y}{\Delta t} & \text{if } (c_0 + U) > 0, \\
0 & \text{if } (c_0 + U) \leq 0.
\end{cases} \tag{2.18}
\]

Here \( U \) is the background flow velocity normal to the boundary. For the alongshore uniform simulation we augment with the background flow on outflow only. Augmenting the velocity from a radiation condition with the advection speed of the flow was introduced by Stevens (1990) for an idealized simulation and employed in a realistic simulation by Stevens (1991). In the implementation described here, the first (outflow) case gives

\[
\varphi_{j}^{n+1} = \varphi_{j-1}^{n},
\]

while the second (inflow) case gives:
So for outflow the variable $\varphi$ is moved across the boundary, whereas for inflow the value is prescribed from a previous time.

2.10 REFERENCES


3 MODELING STUDIES OF THE COASTAL CIRCULATION OFF NORTHERN CALIFORNIA:
STATISTICS AND PATTERNS OF WINTERTIME FLOW

Julie D. Pullen and John S. Allen

To be submitted to Journal of Geophysical Research
3.1 ABSTRACT

We conduct modeling studies of coastal circulation off Northern California in the vicinity of the Eel River (40.6°N, the site of the STRATAFORM marine geology observational program) using a series of nested hydrostatic, primitive equation models. To aid in the understanding of the shelf flow field that receives and transports the sediments during the winter floods, we pursue modeling studies of the 1996–1997 flood season. The basic objectives of our numerical studies are to model the continental shelf and slope flow and to understand the dominant dynamical processes. The validity of the model simulations is assessed by comparison with current measurements from a STRATAFORM shelf tripod outfitted with an acoustic doppler current profiler (ADCP) and from satellite altimetry. We adopt a nesting approach in order to simulate coastal flow on a high-resolution grid of a limited-area domain embedded in a lower-resolution grid of a larger scale domain. For the nested simulations, the outer model is a regional model of the North Pacific Ocean (the Naval Research Laboratory's Pacific West Coast Model, NRL PWC) with approximately 9 km resolution. By implementing a nest of approximately 3-km grid spacing that has high-resolution bathymetry and coastline and realistic river run-off, we simulate the shelf and slope flow surrounding Cape Mendocino during 100 days in winter 1996–1997. The 3-km resolution model exhibits many features of the observed strongly wind-forced shelf flow regime. Good agreement is found between the amplitude and time variability of the 3-km resolution model currents and the observed currents on the shelf. The 3-km resolution model outperforms the 9-km resolution NRL PWC model. Using statistical maps of flow variables, strong alongshore variability in wintertime flow is documented. The role of the major forcing mechanisms in establishing this flow asymmetry between regions north and south of Cape Mendocino is examined us-
ing empirical orthogonal functions. The evolution of a robust anticyclonic eddy over the shelf and slope adjacent to Cape Mendocino is described. The eddy forms when strong poleward winds weaken and reverse direction during winter storms.

3.2 INTRODUCTION

Atmospheric circulation during winter over northern California is dominated by large low-pressure systems with length scales of hundreds of kilometers that pound the coastal region. Maximum wind variability is typically found north of Cape Mendocino, with variability decreasing to the south of the cape (Halliwell and Allen, 1987; Strub et al., 1987). The veering of storm-derived winds is a major forcing mechanism for northern California winter oceanic circulation over the shelf (Largier et al., 1993). In addition to generating significant variability in wind forcing, the storms deliver precipitation to the coastal mountain range. The fresh water is transported rapidly to the coastal ocean by rivers. The resultant river run-off is a source of buoyancy that provides an additional forcing mechanism for the circulation on the northern California shelf.

The winter season of 1996–1997 contained several major storm events. One severe storm produced a massive Eel River flood event with a recurrence time of 80 years (Syvitski and Morehead, 1999). Indeed, winter 1996–1997 contained a confluence of forcing factors that enriched the circulation around Cape Mendocino. The major storms that generated intense winds and caused significant flooding of northern California rivers augmented the complexity of the circulation already complicated by tortuous bathymetry and sharp coastline curvature at Cape Mendocino.
The shelf and slope region surrounding Cape Mendocino was heavily instrumented as part of the Northern California Coastal Circulation Study (NCCCS), 1987–1989 (EG&G, 1991). The goal of that study was to describe seasonal patterns of alongshore variability over the shelf and slope off northern California. The observation period contained one winter (1988–1989). More recently, since 1995 the Strata Formation on Margins (STRATAFORM) observational program has collected data in the region north of Cape Mendocino (Nittrouer, 1999). That program seeks to understand the formation of shelf stratigraphy by deposition of sediments from the Eel River. Sediment deposition is intensified during winter floods, so instrument deployment is focused around these times. Physical measurements include moorings over the shelf and slope outfitted with current meters and temperature/salinity sensors, as well as two tripods on the shelf instrumented with acoustic doppler current profilers (ADCPs). Though much is known about the northern California shelf and slope region from the analysis of observations during spring and summer upwelling — e.g., Largier et al. (1993) and the Coastal Ocean Dynamics Experiment (CODE) as described by Beardsley and Lentz (1987) — winter circulation has only recently received attention (Dever, 1997a; Lentz and Trowbridge, 2000).

Reflecting the focus of observational efforts, spring/summer upwelling conditions off the U.S. West Coast have been modeled at high (less than 5 km) resolution (Oey, 1996; Gan and Allen, 2000). However the winter season has not been the subject of realistic high-resolution modeling studies. In addition, the coastal ocean surrounding Cape Mendocino has not previously been investigated in a modeling study. The numerical study described here constitutes a realistic simulation of winter 1996–1997 flow conditions in the vicinity of Cape Mendocino by including accurate representation of the major forcing mechanisms: wind
stress, river run-off, open-ocean variability, and flow interaction with coastline curvature and bathymetry.

The finite-difference numerical model, domain, forcing, and nesting method are described first. Then statistics over several winter months of the 3-km nested model simulation are analyzed in order to characterize the large alongshore scale (order 100 km) spatial properties of the coastal flow north and south of Cape Mendocino as well as the influence of river run-off on shelf circulation. The nature of fluctuations over the shelf is described by analyzing model results on across-shelf sections. Next we focus on the highly variable region north of Cape Mendocino which was instrumented by the STRATAFORM program. Model time series at a site on the shelf are compared with the observed time series. Dynamical balances are presented for that location. Finally, we identify robust eddy formation generated in response to the passage of intense storms.

3.3 MODEL CONFIGURATION

The Naval Research Laboratory (NRL) has developed a model to reproduce the seasonal circulation off the U.S. West Coast (Clancy et al., 1996). The Pacific West Coast (PWC) model is a version of the stratified hydrostatic primitive equation Princeton Ocean Model (Blumberg and Mellor, 1987). The simulation domain comprises the North Pacific Ocean from 30°N to 49°N and from 135°W to the West Coast (Fig. 3.1). The domain is represented at 1/12° (about 9 km) horizontal resolution and has 30 vertical sigma levels. Navy Digital Bathymetric Data Base (DBDB) 5' (approximately 8 km resolution) bathymetry is used. At the open boundaries the model is coupled by a one-way nest to the six-layer finite-depth 1/6° global Navy Layered Ocean Model (NLOM) (Wallcraft, 1991).
At NRL the PWC model was spun up for two years then forced with Navy Operational Global Atmospheric Prediction System (NOGAPS) (Hogan and Rosmond, 1991) wind stress from 1992 to 1998.

Two one-way nested grids have been configured in the NRL PWC model (Fig. 3.1). The first nested grid extends approximately 200 km offshore (from the coast to 127.1°W) and 475 km alongshore (38.6°N to 43°N) and has about 3 km horizontal resolution. The second nested grid lies within the domain of the first nested grid and extends 275 km alongshore (39.5°N to 42°N) and about 125 km offshore (to 125.9°W) and has approximately 1 km resolution. Like the NRL PWC model, the nested models use 30 vertical sigma levels. High-resolution bathymetry from the Navy database is used on both nested grids. Interpolation to the nested grid boundaries of horizontal velocities in transport form and surface elevation follows the conservative formulation of Clark and Fancy (1984) as described in Chapter 2. The conservative interpolation formulation derived by Clark and Farley (1984) is used in order to maintain consistency of the mass fluxes in the continuity equation between the grids of differing resolution. Other scalar prognostic variables are interpolated linearly. The first nested model was initialized with NRL PWC model fields from October 25, 1996 and subsequently run for over 100 days. The second nested model was used for a 40-day simulation of the major Eel River flood event. Pullen and Allen (2000) analyze the second nested model simulation of the Eel River flood event. The present paper centers on the first nested model (3 km resolution), hereafter referred to as the “nested model” in the analysis sections.

Internal (external) timesteps are 600 s (12 s) for the NRL PWC model; 120 s (4 s) for the first nest; and 24 s (0.8 s) for the second nest. Nested grid boundary values are saved every timestep from the coarser resolution model for all prog-
FIGURE 3.1 Model domain and bathymetry of the Naval Research Laboratory’s Pacific West Coast Model (NRL PWC). Bathymetry contour interval is 500 m. Boundaries of the nested models are indicated.
FIGURE 3.2 NOGAPS model wind stress mean and rms vector amplitude over the 100-day simulation on the NRL PWC domain.
nostic variables. In order to update the boundary of the nested grids, prognostic variables are interpolated linearly in time to provide values every timestep. More details of the nesting implementation are described in Chapter 2.

3.4 MODEL FORCING

NOGAPS model wind stress variability over the domain of the NRL PWC model for the 100-day simulation (Fig. 3.2) is strongest north of 39°N and weakens to the south. Mean wind stress is oriented alongshore to the north of Cape Mendocino, while to the south mean wind stress has an onshore orientation. NOGAPS model wind stress at NDBC Buoy 46030 (40.4°N, 124.5°W – hereafter called Buoy #30), located adjacent to Cape Mendocino, agrees well with stress calculated from winds measured at that buoy (Fig. 3.3) (correlation coefficient = 0.92; NOGAPS (observed) mean = 0.65 (0.46) dyne cm⁻²; NOGAPS (observed) standard deviation = 1.73 (1.30) dyne cm⁻²). Wind stress from the observed wind speeds was computed using the method of Large and Pond (1981). NOGAPS wind stress was used to force the NRL PWC model as well as the two nested models within. Two major storm events on December 8–17 and December 29, 1996 – January 6, 1997 are evident in the wind record. Strong poleward (downwelling-favorable) winds give way to equatorward (upwelling-favorable) winds as the midlatitude cyclones propagate eastward. Later analysis will concentrate on these two storm passages.

River discharge values recorded hourly by the United States Geological Survey (USGS) at stream-gaging stations and archived by the California Department of Water Resources (CDWR) were low-pass filtered (Butterworth filter with cut-off period of 3 hours) then introduced into the NRL PWC model and interior nested
FIGURE 3.3 Model forcing for the NRL PWC and nested models for the 100-day simulation. NOGAPS model wind stress at the location of Buoy #30 (40.4°N, 124.5°W) is shown along with wind stress calculated from observed wind speed at the buoy. The major axis direction has been chosen. River gauge discharge data for the Klamath and Eel Rivers is shown in the bottom two panels.
models using the method of Kourafalou et al. (1996) (see Appendix). Discharge measured at the Turwar station was used for the Klamath River. Discharges measured at the Scotia (Eel River) and Bridgeville (Van Duzen River) stations were combined to give discharge near the mouth of the Eel River (Wheatcroft et al., 1997). Discharge of the Rogue River was taken from climatology (Clancy et al., 1996). River run-off from both the Eel and Klamath Rivers (Fig. 3.3) peaks as the trailing end of the storms pass over the region around December 10 and January 2. The nested model simulation was repeated with no river run-off in order to differentiate the impact of freshwater buoyancy forcing from wind forcing on the northern California coastal circulation.

Observations from the NCCCS study suggest that mesoscale features from the open ocean can influence the shelf and slope circulation around Cape Mendocino (Largier et al., 1993). Open ocean features including jet meanders and eddies that impinge on the shelf and slope may be in the initialized fields from the coarser resolution model and may be advected as the simulation progresses by a nested modeling approach.

The interaction of flow with the convoluted bathymetry around Cape Mendocino provides a final important forcing mechanism. Navy DBDB 0.5' (approximately 0.5 km) and 1' (approximately 1.5 km) digitized bathymetry is combined to create high-resolution bathymetry for the domain inside the first nested grid boundary. Bathymetry on the nested grid is smoothed and blended with the coarser resolution model bathymetry over six grid cells for consistency across the nested grid boundary. Immediately off Cape Mendocino as the coastline projects westward, the bathymetry becomes highly contorted, with the Mendocino Escarpment (Fig. 3.1) dropping sharply to the north. Furthermore, several deep submarine canyons gouge the slope area adjacent to the Cape Mendocino coast.
There is a broad shelf north of Cape Mendocino while to the south the shelf is narrow. The high-resolution bathymetry and coastline used on the nested grid captures the important features of the coastline and underwater terrain (Fig. 3.4). (See Pullen and Allen (2000) for the bathymetry used on the second nested grid.)

3.5 CIRCULATION STATISTICS

This section gives a statistical overview of the model results of the 100-day nested model (3 km resolution) simulation. Because most of the forcing mechanisms described in the previous section possess strong alongshore variability, it is natural to expect the coastal flow around Cape Mendocino to mirror this spatial inhomogeneity. This section focuses on comparing and contrasting statistical quantities at different locations in order to identify and separate the effects of the dominant physical forcings. All variables in this paper are filtered with a moving average over the 18-hour inertial period, unless otherwise stated.

As a point of departure for the analysis, current meter measurements from the NCCCS observational program are examined. The same 100-day period corresponding to the model simulation is utilized (November 5, 1996 - February 13, 1997), however the observations are for the year 1988–1989. (It was not practical to run the nested model for the 1988–1989 period of the NCCCS experiment because the NRL PWC simulation was conducted for 1992–1998 and NOGAPS wind forcing is not readily available for 1988–1989.) The historical NCCCS current meter data on the Klamath (KLA) and Vizcaino (VIZ) lines shown in Fig. 3.4 and summarized in Table 3.1 reveals that the principal axis ellipses of standard deviation of velocity at 10 m depth are more polarized alongshore on the VIZ line (south of Cape Mendocino) compared with the KLA line (north of Cape Men-
TABLE 3.1 Statistics from the NCCCS observations of currents at 10 m depth (cm s\(^{-1}\)) and wind stress (dyne cm\(^{-2}\)) from November 5, 1988 to February 13, 1989.

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<th>#30 (wind stress)</th>
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<th>direction</th>
<th>standard dev. magnitude</th>
<th>standard dev. direction</th>
</tr>
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<td>(mean)</td>
<td>(major axis)</td>
<td>(minor axis)</td>
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</tr>
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</tr>
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<td>S</td>
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<td>7.54</td>
</tr>
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<td>8.20</td>
</tr>
<tr>
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</tr>
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<td>2.34</td>
<td>SW</td>
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<td>13.31</td>
<td>5.03</td>
</tr>
<tr>
<td></td>
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<td>direction</td>
<td>standard dev.</td>
<td>standard dev.</td>
</tr>
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<td>(mean)</td>
<td>(major axis)</td>
<td>(minor axis)</td>
</tr>
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</tr>
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<td>SE</td>
<td>11.56</td>
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</tr>
<tr>
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<td>SE</td>
<td>16.31</td>
<td>3.44</td>
</tr>
<tr>
<td>VIZ 130 m</td>
<td>5.83</td>
<td>SE</td>
<td>20.89</td>
<td>4.67</td>
</tr>
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</table>

**TABLE 3.2** Statistics from the nested model simulation of currents at 10 m depth (cm s\(^{-1}\)) and wind stress (dyne cm\(^{-2}\)) from November 5, 1996 to February 13, 1997.
The shelf is narrower at the VIZ line than at the KLA line, which can account for the observed difference in polarization. Mean flow over the shelf and slope at the KLA and VIZ lines is generally equatorward despite weak poleward mean winds. The wind forcing during winter 1988–1989 at NDBC Buoy #30 possessed a large number of equatorward wind events, leading NCCCS investigators to categorize winter 1988–1989 as anomalous (EG&G, 1991). By contrast, winter 1996–1997 displayed large mean poleward winds and higher wind variability (Fig. 3.4). Table 3.2 contains the statistics from the model simulation. Model currents at 10 m depth from the present simulation (November 5, 1996–February 13, 1997) also show stronger polarization of flow in the alongshore direction off VIZ compared to KLA. Model flow is equatorward in the mean except over the shelf on the 60 m isobath off KLA where it is poleward. The “no river” simulation also had poleward flow on the 60 m isobath at KLA. Model variability is stronger at all locations than during the NCCCS observational period.

Time-mean surface velocity and rms amplitude of surface velocity for the 100-day simulation with and without rivers (Fig. 3.5) indicates that the buoyancy forcing from the Eel and Klamath Rivers enhances the fluctuations in the surface current over the shelf. The effect of river run-off contributes up to 12 cm s⁻¹ to the surface velocity variability over the shelf. This contribution is not distributed uniformly alongshore. Patches of large fluctuations occurring without river run-off north of Point St. George and between the Eel River and Trinidad Head are augmented when river run-off is included. This enhancement of flow variability due to buoyancy forcing is confined to the areas north of Cape Mendocino and is superimposed on the alongshore asymmetry in the current fluctuations that exists between the region north and south of Cape Mendocino in the absence of river run-off. Mean surface velocity diverges at Cape Mendocino and flows poleward
FIGURE 3.4 Northern California Coastal Circulation Study (NCCCS) and nested model 10 m velocity and wind stress principal axis ellipses of standard deviation. Standard deviations correspond to the major axes of the ellipses. Bathymetry contours on both plots represent the actual bathymetry used on the nested model domain. The model domain is truncated to the west.
to the north of the cape and equatorward to the south. This divergence pattern with equatorward flow south of Cape Mendocino also occurs in the NRL PWC mean surface velocity field and may be associated with the large-scale negative wind stress curl south of Cape Mendocino (Fig. 3.2).

Depth-averaged time-mean velocities (Fig. 3.6) are muted over the shelf and their direction along the coast is more variable than was the surface velocity time-mean. The asymmetry in variability between regions north and south of Cape Mendocino and the enhancement of variability in the same two areas described above is readily apparent. Here, however, there is no discernible difference between variability in the flow with and without rivers (not shown). Mean surface elevation (Fig. 3.6) has anticyclonic circulation to the north and south of Cape Mendocino. Variability of sea surface elevation is greatest north of Trinidad Head.

The differences in flow variability north and south of Cape Mendocino, as well as differences due to river run-off, can be further assessed by an examination of model transects of time-means, standard deviations and empirical orthogonal functions (EOFs) along east-west (across-shore) sections at VIZ and S60 (Figs. 3.7, 3.8, 3.9). The plots involving alongshore velocity in Fig. 3.7 highlight similarities and differences north and south of Cape Mendocino and are calculated from the 100-day simulation without rivers. A poleward jet over the mid-shelf dominates the mean at S60; elsewhere the flow is equatorward. At the VIZ section the mean velocity is equatorward. EOFs partition the variance in a data set into ranked orthogonal spatial modes that cumulatively account for the variability in the data. EOFs, by identifying coherent spatial patterns whose amplitude is modulated in time, permit the association of these patterns (modes) with physical forcings that have similar modulations in time. Based on the structure and amplitude of the EOFs, fluctuations in the velocity display
FIGURE 3.5 Nested model surface velocity vector mean and rms vector amplitude over the 100-day simulation, with rivers and without rivers. Means are denoted with arrows while rms amplitude is color contoured. Every second vector is shown in the along-shore and across-shore direction.
FIGURE 3.6 Nested model depth-averaged velocity vector mean and rms vector amplitude over the 100-day simulation with rivers. Means are denoted with arrows; rms amplitude is color contoured. Every second vector is shown in the along-shore and across-shore direction. Also shown is the mean (black contours) and standard deviation (color contours) of surface elevation over the 100-day simulation with rivers.
a similar pattern at the two transects. The majority of the variance is associated with coastal jet-like motions, concentrated in the upper 40 m, of northward fluctuations during poleward winds and southward fluctuations during equator-wind winds. This is consistent with idealized shelf model response to upwelling and downwelling-favorable wind forcing (Allen and Newberger, 1996; Allen et al., 1995).

Plots in Fig. 3.8 document the influence of river discharge on the variability north of Cape Mendocino. Compared to the no river case (Fig. 3.7), mean poleward flow in the jet is slightly stronger when rivers are present. Variability occurs mainly in the upper 5-10 m. This is shallower than the fluctuations that occur without rivers. Mean density is distinguished by light Eel River water in the upper 5-10 m. Density fluctuations are concentrated near the surface and occur when water lighter than the mean spreads out over the shelf during the Eel River flood events.

To investigate further the differences in current variability near the surface with and without rivers we look at the turbulent kinetic energy \( \frac{1}{2}q^2 \) from the Mellor and Yamada (1982) level 2.5 turbulence submodel along the S60 transect, with and without rivers (Fig. 3.9). Without rivers, turbulent kinetic energy variability is large in the upper 25 m from mid-shelf to offshore. With rivers, fluctuations in \( \frac{1}{2}q^2 \) are most intense near the base of the strong jet, suggesting the presence of shear-induced turbulence. The \( \frac{1}{2}q^2 \) amplitude time series tracks wind stress so that turbulent kinetic energy increases during strong wind events. The strong stratification caused by river run-off may confine intense \( \frac{1}{2}q^2 \) values over the midshelf to within about 10 m of the surface. Though \( \frac{1}{2}q^2 \) means and standard deviations are larger in this near-surface region for the “rivers” case compared with the “no rivers” case, nonetheless, over midshelf the “no rivers”
case has large $\frac{1}{2}q^2$ values that extend more deeply vertically by about 15 m than those of the "rivers" case.

We have assessed the structure of means, standard deviations and dominant EOFs at isolated transects. To conclude this section we evaluate the degree of communication between regions north and south of Cape Mendocino. Fig. 3.10 shows the alongshore space-lagged velocity correlation coefficients relative to velocity at KLA and VIZ at 50 m depth over the 100 m isobath. At each location along the coast the velocity is rotated into the principal axes of the depth-averaged velocity. Correlation coefficients are calculated for both the major and minor axis components. Minor axis correlation scales are on the order of 10 km. Short alongshore correlation scales of the across-shore (minor axis) velocity components are typical features of coastal flow regimes off the northern U.S. west coast (Kundu and Allen, 1976; Dever 1997b). Major axis correlation lengths extend around Cape Mendocino. However the magnitude of the correlations falls rounding Cape Mendocino. For the correlations anchored at KLA, magnitudes are lower to the south of Cape Mendocino but remain high to the north. For the correlations anchored at VIZ, magnitudes drop off to the north and south.

3.6 TIME SERIES

Modeled and observed major axis velocity time series at three depths at the S60 ADCP site (40.89°N, 124.25°W) are compared in Fig. 3.11. The S60 ADCP (RDI-workhorse) observations are described in Ogston et al. (2000). The correlation coefficients are considerably higher for the nested high-resolution model (3 km) relative to those of the NRL PWC model (9 km) (Table 3.3). Correlations remain high for the nested model throughout the water column; whereas correla-
FIGURE 3.7 Model mean (contour lines), standard deviation (color shading) and 1st EOF of north-south velocity at transects S60 and VIZ for the nested simulation without rivers. The percentage of variance explained by the 1st EOF is indicated. The amplitude time series of the 1st EOF of velocity is shown at the bottom. Wind stress at Cape Mendocino (Buoy #30) is included for comparison.
FIGURE 3.8 As in Fig. 3.7 but at S60 for simulation with rivers. Top panel is north-south velocity ($v$); bottom panel is potential density ($\sigma_\theta$). Corresponding amplitude time series are shown below.
Turbulent Kinetic Energy

FIGURE 3.9 As in Fig. 3.7 but for turbulent kinetic energy, $\frac{1}{2}q^2$, at S60 with and without rivers. Turbulent kinetic energy has been multiplied by $10^2$. 
<table>
<thead>
<tr>
<th>major axis velocity (cm s(^{-1}))</th>
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<th>mean</th>
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</tr>
</thead>
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<td>22.12</td>
</tr>
<tr>
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<td>0.81</td>
<td>3.09</td>
<td>21.45</td>
</tr>
<tr>
<td>Modeled (9 km) 10 m</td>
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<td>5.83</td>
<td>16.49</td>
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<tr>
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<td>17.91</td>
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<td>16.15</td>
</tr>
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<td>Modeled (9 km) 50 m</td>
<td>0.32</td>
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<td>13.85</td>
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TABLE 3.3 Model and observed time series statistics at STRATAFORM S60 ADCP site. Observed data used courtesy of A. Ogston.

...tions drop off sharply with depth for the NRL PWC model. Observed time-mean flow is equatorward at all depths. Neither model reproduces the mean flow. The second nested model (1 km), however, does reproduce the observed time-mean vertical current structure (Pullen and Allen, 2000). Observed standard deviations are high near the surface and are lower near the bottom. The magnitudes of the standard deviations from the nested model (3 km) are in reasonable qualitative agreement with the observed values and replicate the reduction with depth. The NRL PWC model has standard deviations that are smaller in magnitude at 30 m and above and are more uniform with depth.

Terms in the nested model depth-integrated alongshore momentum balance (Blumberg and Mellor, 1987)
FIGURE 3.10 Zero-lagged correlation coefficients for principal axis velocity following the 100 m isobath at 50 m depth. The correlations are relative to the velocity at KLA (top) and VIZ (bottom).
FIGURE 3.11 Modeled and observed major axis velocity at the S60 ADCP location on the shelf at three depths over the 100-day record. Velocities are rotated into the principal axes of the depth-averaged current. Observations are shown here courtesy of A. Ogston.
\[
\frac{\partial V D}{\partial t} + \frac{\partial U V D}{\partial x} + \frac{\partial V^2 D}{\partial y} - F_y - G_y + f U D + \frac{\partial}{\partial y} \tau_{yD} - \tau_{yh} + \frac{g D \partial \eta}{\rho_0} + \frac{g D}{\rho_0} \int_{-1}^{0} \int_{\sigma}^{0} \left( D \frac{\partial \rho'}{\partial y} - \frac{\partial D}{\partial y} \sigma \frac{\partial \rho'}{\partial \sigma} \right) d\sigma' d\sigma = 0
\]  

(3.1)

during the two major storm events at the S60 site are shown in Fig. 3.12, after division by the depth D, along with the depth-averaged current. The terms are: (1) acceleration, (2) advection, (3) Coriolis force, (4) surface stress, (5) bottom stress, and (6) pressure gradient. The alongshore balance above is obtained by rotating the original x (east-west) and y (north-south) equation into the principal axis orientation of the depth-averaged current.

The time-dependent term balance and currents demonstrate a generic response to passing storms. In particular, the surface stress exerted by the winds accelerates a poleward current. Bottom friction increases to retard the accelerating current. Meanwhile, an equatorward pressure gradient that primarily balances the wind-forced offshore flow below the surface layer develops. After the relaxation of the surface stress, the wind-forced across-shelf flow decreases so that it is no longer large enough to balance the pressure gradient. As a result, the surviving pressure gradient accelerates the alongshore current southward and leads to current reversals on December 14 and January 3. The major axis depth-averaged currents and the important terms in the momentum balance have larger magnitudes in response to the second (stronger) storm. The minor axis momentum balance is predominantly geostrophic and is not shown.

The development of the southward pressure gradient force is shown in maps of surface elevation (Fig. 3.13). After December 28, as poleward winds grow strong, the surface elevation develops alongshore variations. In particular, the regions south of Trinidad Head (41.1°N), where S60 is situated, and south of Point George
(41.7°N) experience a sea level set-up which establishes an alongshore pressure gradient.

3.7 CIRCULATION PATTERNS

Nested model sea surface elevation during the second major storm is shown in Fig. 3.14 for the simulation with and without rivers. Winds are strongly poleward on December 31. Sea level set-up against the coast, emblematic of downwelling, is evident at these times. Alongshore spatial inhomogeneities in sea level elevation are present due both to coastline irregularity and rivers. River run-off creates a buoyancy current that augments the sea level set-up and enhances the coastal downwelling jet (Fong, 1998).

On January 6, wind stress has reversed to equatorward and sea level elevation domes to form an anticyclonic eddy centered near 40.75°N. The strength of the elevation "high" is increased by river run-off, but the eddy is present even without river run-off. By contrast, the high sea level cell adjacent to the Klamath River is absent when river run-off is not included. The same pattern of eddy occurrence with and without rivers appears during the first storm; however the magnitude of sea level elevation is reduced.

Fortuitously, an ERS-2 satellite track crossed the waters near Cape Mendocino on January 6. Topex/ERS-2 satellite altimetry confirms the presence of a clockwise (anticyclonic) eddy adjacent to Cape Mendocino (Fig. 3.15) on January 6. The broad poleward flow offshore of the remotely sensed eddy is also found in the model simulation. Furthermore, the anticyclonic circulation south of Cape Mendocino in the satellite field is also indicated in the model sea surface height field.
FIGURE 3.12 Nested model depth-averaged current and depth-integrated major axis (alongshore) momentum balance at S60 during two storm events. The terms in the momentum equation (3.1) have been divided by the depth $D$ and multiplied by $10^7$. 
FIGURE 3.13 Evolution of sea surface elevation during the early stages of the second severe storm. Units are cm.
FIGURE 3.14 Sea surface elevation at two times during the second major storm, with rivers and without rivers. Units are cm. The east-west line at 40.75°N marked on the January 6, 1997 “Rivers” panel is the location of the east-west sections plotted in Fig. 3.17.
FIGURE 3.15 TOPEX/ERS-2 altimeter snapshot of height field anomalies. Units are cm. Processing of satellite data was done by C. James.
FIGURE 3.16 Maps of depth-averaged velocity north of Cape Mendocino. Wind stress at Buoy #30 is indicated with a thick arrow. The eddy formation adjacent to the Eel and Klamath Rivers is shown.
Velocity

January 2, 1997

January 4, 1997

January 6, 1997

FIGURE 3.17 East-west section of north-south velocity through the eddy pictured in Fig. 3.14. Units are cm s\(^{-1}\). The model simulation with rivers is shown. The sections extend approximately 60 km offshore.
The evolution of the eddy near the end of the second storm is traced in maps of depth-averaged velocity (Fig. 3.16). The reversal of the coastal current from poleward (January 2) to equatorward (January 5) is accompanied by a complicated pattern of eddies and meanders. Notably, the eddies adjacent to the Eel and Klamath Rivers start to form January 3. From the behavior of the depth-averaged velocity vectors in the vicinity of the Eel River, it appears that the shelf and slope bottom topographic variations associated with the Eel River canyon just south of about 40.6°N play a major role in the formation of the eddy in this region.

An east-west section of velocity cutting through the eddy (See Fig. 3.14 for location of the section) extending about 60 km offshore shows the depth structure of the eddy (Fig. 3.17). The eddy reaches deeper than 500 m and has a subsurface maximum of 25 cm s\(^{-1}\) at about 100 m depth. The equatorward limb of the eddy forms over the shelf by January 6. At later times the eddy moves offshore and decays as upwelling-favorable winds influence the region (not shown). The eddy has a similar depth structure during the first storm, though the velocities are about 5-10 cm s\(^{-1}\) weaker.

The eddy circulation associated with the Klamath River to the north has very different characteristics. That eddy occurs only when rivers are present in the simulation and is situated over the shelf. Velocities in that eddy are surface-intensified during both storms.

The signature of the eddy near the Eel River comes through in the mean depth-averaged current (Fig. 3.6), as well as the mean surface velocity (Fig. 3.5). Maps of the first and second mean product EOFs of surface currents calculated with the mean values retained (Fig. 3.18) and corresponding amplitude time series (Fig. 3.20) identify the eddy as part of a mode 2 (explaining approximately 14% of
the mean product) response to the relaxations/reversals of poleward winds. This second mode is characterized during the wind relaxation by mostly equatorward flow over the shelf. In this mode, an anticyclonic eddy is also found to the south, in the fold of Cape Mendocino. The relatively large amplitudes and large fraction of local mean product explained in the near-coastal surface currents south of Cape Mendocino indicate that this mode, whose variability is separate from that of the mode 1 large scale wind-forced response, makes a dominant contribution to the mean southward currents south of Cape Mendocino (Fig. 3.5).

The first mode EOF (explaining about 58% of the mean product) represents a coastal current that has a larger magnitude and represents a larger fraction of the local mean product north of Cape Mendocino. The variability of the first mode is clearly related to that of the wind stress (Fig. 3.20). The maximum time-lagged correlation coefficient is 0.89 at a lag of 24 hours. This mode represents a large alongshore-scale coastal circulation response to wind forcing with greatest strength north of Cape Mendocino. The portion of the surface velocity mean product attributable to the other modes drops off sharply; mode 3 contributes 7%. Thus, two modes explain the majority of the surface velocity mean product and characterize the coherent coastal surface current response to storm events.

The depth-averaged velocity mode 1 EOF (Fig. 3.19), calculated as is usual with the mean values removed, represents a large-scale coastal current fluctuation that is strongest north of Cape Mendocino, but that is also appreciable, both in relative amplitude and in fraction of local variance represented, south of Cape Mendocino. The amplitude time series of this mode is highly correlated with the wind stress \( r = 0.85 \) at a lag of 36 hours. Thus, this mode represents a large-scale coherent coastal current response to wind that extends from about 39°N to north of 43°N. The extension of this coherent wind-driven response south of Cape
Mendocino was masked in the surface current mean product EOFs in Fig. 3.18 by the unrelated variability of the mode 2 coastal surface currents south of Cape Mendocino.

3.8 DISCUSSION AND SUMMARY

We have conducted the first high-resolution numerical study of winter circulation around Cape Mendocino. We have examined the contributions of major forcing mechanisms to the circulation in the region using a statistical approach. In the absence of river forcing the asymmetry in flow variability north and south of Cape Mendocino, with the flow exhibiting more energetic behavior to the north, is likely related to the coastline and bathymetry variations, since regions north and south of Cape Mendocino show a similar response pattern to the wind stress.

Primarily associated with the flood events, river discharge bathes the near-surface (upper 5–10 m) waters of the shelf, with light water enhancing the stratification near the surface. River run-off increases the magnitude of velocity fluctuations in the coastal jet.

Major and minor axis 50 m alongshore velocity correlations along the 100 m isobath were computed at points north and south of Cape Mendocino. Minor axis (across-shore) correlation scales are about 10 km. Major axis (alongshore) correlation scales are greater than 100 km over the broad shelf region to the north of Cape Mendocino, and shorter on the narrower shelf to the south of Cape Mendocino. Alongshore velocity correlations are reduced to the south of Cape Mendocino.
FIGURE 3.18 Map of dominant surface velocity mean product EOFs (arrows). Modes 1 and 2 are shown. Percent mean product explained by each mode is given in the title. Color contours show the amount of total mean product at each site explained by the mode. The simulation with rivers is used.
FIGURE 3.19 Map of mode 1 depth-averaged velocity EOF (arrows). Percent variance explained by the mode is given in the title. Color contours show the amount of total variance at each site explained by the mode. The simulation with rivers is used.
FIGURE 3.20 Amplitude time series for modes 1 and 2 of the mean product EOFs of surface velocity shown in Fig. 3.18 (top) and amplitude time series for mode 1 of the EOF of depth-averaged velocity shown in Fig. 3.19 (bottom). NOGAPS wind stress at Cape Mendocino (Buoy #30) is also shown.
Good agreement is found between the amplitude and time variability of the nested model currents and the observed ADCP currents at three depths at the S60 location during the 100-day simulation. The nested model (3 km) performed better than the NRL PWC (9 km) model, especially at deeper depths. At the shelf site, terms in the depth-integrated major axis momentum balance displayed a common time evolution to passing storms. This response includes the establishment of an alongshore pressure gradient force that helps initiate and sustain a current reversal over the shelf. This current reversal at the shelf site develops into the shoreward arm of an anticyclonic eddy. The eddy occurrence was verified by satellite altimetry. In addition, the eddy was described as a local generic response to traveling atmospheric cyclones through an EOF analysis of velocity. The interaction of the Eel River plume with the eddy and the implications of eddy circulation for sediment transport on the shelf are addressed in Pullen and Allen (2000).

3.9 ACKNOWLEDGMENTS

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3.10 RIVER INPUT

River run-off from the Eel and Klamath rivers is discharged into all three model domains using the method described by Kourafalou et al. (1996). The modifications to the governing equations at the location of the river input are reviewed here.

The vertically integrated continuity equation becomes

$$\frac{\partial \eta}{\partial t} + \frac{\partial \bar{U} D}{\partial x} + \frac{\partial \bar{V} D}{\partial y} = \bar{q}.$$  

Here $\bar{q} = Q/\Delta x \Delta y$ where $Q$ is the river discharge (in m$^3$ s$^{-1}$) and $\Delta x, \Delta y$ are the horizontal grid spacing.

The three dimensional continuity equation becomes

$$\frac{\partial \eta}{\partial t} + \frac{\partial \omega}{\partial \sigma} + \frac{\partial U D}{\partial x} + \frac{\partial V D}{\partial y} = q'(k),$$

where $q'(k)$ must satisfy the constraint

$$\sum_{k=1}^{K} q'(k) \Delta \sigma(k) = \bar{q}.$$  

Here $k$ is an index for the vertical sigma ($\sigma$) level, $K$ is the number of $\sigma$ levels, and $\Delta \sigma(k)$ is the spacing between adjacent $\sigma$ levels: $\sigma(k) - \sigma(k+1)$. Let $P(k)$ be the fraction of discharge, $\bar{q}$, put in at $\sigma$ level $k$. Then we find the value of $q'(k)$ by:

$$q'(k) = \frac{P(k) \bar{q}}{\Delta \sigma(k)},$$

where

$$\sum_{k=1}^{K} P(k) = 1.$$
If the river is input uniformly with depth then $P(k) = \Delta \sigma(k)$. Whereas if the river is input only in the top layer, as first described and implemented by Kourafalou et al. (1996) and used in this paper, then $P(1) = 1$ and $P(k) = 0$ for $k \neq 1$.

The salinity equation becomes

$$\frac{\partial S}{\partial t} + \frac{\partial S U}{\partial x} + \frac{\partial S V}{\partial y} + \frac{\partial S \omega}{\partial \sigma} = q'(k) S_{\text{river}} + \frac{\partial}{\partial \sigma} \left[ \frac{K_H}{D} \frac{\partial S}{\partial \sigma} \right] + F_S,$$

while the temperature equation becomes

$$\frac{\partial T}{\partial t} + \frac{\partial T U}{\partial x} + \frac{\partial T V}{\partial y} + \frac{\partial T \omega}{\partial \sigma} = q'(k) T_{\text{river}} + \frac{\partial}{\partial \sigma} \left[ \frac{K_H}{D} \frac{\partial T}{\partial \sigma} \right] + F_T,$$

where $S_{\text{river}}$ and $T_{\text{river}}$ are the salinity and temperature, respectively, of the river discharge. In the application described in this paper, $S_{\text{river}} = 0$ and $T_{\text{river}} = 11.5^\circ C$ on all model domains. In the first nested model the river discharge is put into the top layer at the innermost grid cell of a two-grid-cell-long inlet channel of one-grid-cell width.

### 3.11 REFERENCES


4 MODELING STUDIES OF THE COASTAL CIRCULATION OFF NORTHERN CALIFORNIA: SHELF RESPONSE TO A MAJOR EEL RIVER FLOOD EVENT

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4.1 ABSTRACT

The region surrounding Cape Mendocino is the subject of a high-resolution multiply-nested modeling study. The highest resolution model (1 km) is used for a 40-day simulation containing the major Eel River flood event of December 30, 1996 – January 3, 1997. The complex forcing due to rapidly shifting storm-driven winds, flood-induced river discharge and rugged bathymetry and coastline generate energetic coastal circulation off northern California during this time. The Eel River plume mixes significantly in the down-stream direction and influences the velocity structure of the shelf flow. Coastline irregularities are important in generating an eddy over the Eel River margin through the establishment of an alongshore pressure gradient. Time series and statistics of the modeled alongshore velocity field show good agreement with observed values at the S60 and G65 ADCP sites.

4.2 INTRODUCTION

The northern California shelf region experienced several severe storms in winter 1996-1997. The discharge from precipitation-laden rivers caused major flooding in early January 1997 throughout the region, during which time the Eel River (40.6°N, 124.3°W) experienced an 80-year flood event (Syvitski and Morehead, 1999). The Strata Formation on Margins (STRATAFORM) observational program instrumented the region north of the Eel River during the flood season. The focus of that program is to understand the transport and accumulation of sediments on continental margins (Nittrouer, 1999). Geyer et al. (2000) describe the evolution of a plume formed by Eel River outflow during the flood. They discuss characteristics of the plume, as well as construct
a budget for the sediment carried by the plume. The flood was caused by a passing cyclonic storm system and as a consequence, the coastal region was subjected to strong poleward (downwelling-favorable) winds followed by equatorward (upwelling-favorable) winds. The plume was confined close to the coast and flowed poleward during downwelling-favorable winds. During the subsequent upwelling-favorable winds the plume expanded offshore and thinned vertically. These two plume response patterns to opposing wind conditions have also been documented in observational studies by Rennie et al. (1999) and Münchow and Garvine (1993). As the winds transitioned to upwelling-favorable, the suspended sediment concentration in the plume diminished substantially, indicating sediment fall-out from the plume (Hill et al., 2000). The analysis by Geyer et al. (2000) considered sediment transport by the 5–10 m thick Eel River plume. In particular, their analysis did not directly account for low-frequency shelf circulation that could play a role in determining sediment fate once it sinks out of the plume (Ogston et al., 2000).

Recognized as a source of buoyancy and momentum for the coastal ocean, river run-off has been included in idealized model configurations. Chao (1987) used a primitive equation model and described effects of wind on buoyancy currents. Upwelling winds modified the structure of the current as described above. Fong (1998) extended the work of Chao (1987) by analyzing dynamic river plume responses to upwelling and downwelling winds over a straight coastline. Fong (1998) documented momentum balances and characterized mixing in the river plume for an alongshore-averaged two-dimensional section. Garvine (1999) investigated the dependence of three-dimensional plume characteristics on model parameters.
FIGURE 4.1 Domain and bathymetry of the second nested model (1-km resolution). The locations of STRATAFORM data sites G65, G60 and S60 on the shelf are indicated with crosses. The National Data Buoy Center (NDBC) Buoy 46030 (designated Buoy #30) located off Cape Mendocino is marked with a circled cross.
FIGURE 4.2 Model forcing for the NRL PWC and nested models for the 40-day simulation. Model (NOGAPS) wind stress at the location of Buoy #30 (40.4°N, 124.5°W), about 30 km south of the Eel River mouth, is shown and compared with wind stress calculated from Buoy #30 measured wind speeds (Chapter 3). USGS river gauge discharge values for the Klamath (Turwar station) and Eel Rivers (Scotia and Bridgeville stations combined) are also displayed.
Recent modeling studies have incorporated realistic forcing, including spatially and temporally varying winds and river run-off (e.g., Kourafalou, 1999; Masson and Cummins, 1999). They demonstrate the role of alongshore variability in the shelf dynamics and compare model results with observations.

The present study reports the results of a 40-day simulation focused on the major Eel River flood event of winter 1996–1997 using a high-resolution nested model with approximately 1-km resolution ($\Delta y=1.0$ km, $\Delta x=1.0$ to 1.1 km). We nested a series of local high-resolution models (3-km and 1-km resolution, respectively) within a regional model of the North Pacific Ocean (9-km resolution). The Naval Research Laboratory's Pacific West Coast Model (NRL PWC) was used as the regional model (Clancy et al., 1996). The NRL PWC is an implementation of the Princeton Ocean Model (POM), as are all interior nested models. All models utilize 30 vertical sigma levels. POM is a sigma-coordinate primitive equation model with an embedded turbulence scheme (Blumberg and Mellor, 1987). Chapter 2 contains the details of the nesting approach. In Chapter 3, the first nested model (3-km resolution) simulation of 100 days in winter 1996–1997 is used to describe the circulation with and without river run-off. Strong variability and complex flow is characteristic of the northern Eel River margin even in the absence of river run-off. Chapter 3 reports a robust anticyclonic eddy centered off the Eel River mouth that occurred in response to the transient wind forcing imposed by passing storms. This eddy is strengthened by river discharge and is a circulation response to the shift of winds from strongly poleward to weakly equatorward. In addition, Chapter 3 demonstrates the prevalence of a generic dynamic shelf response to storm events using terms in the depth-integrated along-shore momentum equation. The role of a southward pressure gradient force in
assisting current reversal from northward to southward over the shelf is also emphasized in Chapter 3.

The present chapter will describe the second nested model simulation (1-km resolution). The goal of this work is to characterize the shelf circulation during a 40-day time period surrounding the Eel River flood event, to compare model results with observations during that time period, and to document the Eel River plume interaction with the background shelf flow.

4.3 MODEL FORCING

The model extends 275 km alongshore and about 125 km offshore (Fig. 4.1). Bathymetry was generated from the high-resolution Navy DBDB0.5' and DBDB1' (approximately 0.5 km and 1.5 km, respectively) digitized database. The precipitous shelf and slope terrain of the Mendocino Escarpment is well-resolved. Important coastline irregularities, labeled in Fig. 4.1, are also well-represented by the model resolution.

Wind and river run-off are significant forcing mechanisms in the winter season and are represented realistically in the model simulation (Fig. 4.2). Navy Operational Global Atmospheric Prediction System (NOGAPS) model wind stresses are used to force the NRL PWC model. Over the 40-day period of simulation the NOGAPS wind stresses agree well with stresses calculated from observed winds at Buoy 46030, hereafter designated Buoy #30, adjacent to Cape Mendocino. The correlation coefficient is 0.95, while NOGAPS (observed) mean $\mu = 0.99 \ (0.70) \ \text{dyne cm}^{-2}$, and NOGAPS (observed) standard deviation $\sigma = 1.89 \ (1.49) \ \text{dyne cm}^{-2}$. Therefore NOGAPS winds are also used to force the nested models.
FIGURE 4.3 Observed and modeled vertical structure of the time-mean (solid) and standard deviation (dashed) of major axis (alongshore) velocity at S60 and G65. Statistics are computed over a common overlapping ADCP deployment period of 32 days. The ADCP data are shown courtesy of C. Friedrichs (G65) and A. Ogston (S60).
Low-pass filtered hourly river discharge of the Eel and Klamath rivers from USGS river gauge stations (Fig. 4.2) is introduced into all model domains by the technique of Kourafalou et al. (1996). The river inlet is three grid cells wide to match the inlet width on the 3-km domain. The inlet length is five grid cells. The river mass flux is input at the top vertical level at the innermost grid cells at the head of the inlet. (See Chapter 3 or details of the river implementation.)

After January 2, winds reverse to upwelling-favorable as discharge from the Eel River decreases. By contrast, discharge from the Klamath River remains at peak during this wind relaxation. The abrupt switch from downwelling to upwelling-favorable winds as the storm passes and the timing of this wind reversal relative to the unsteady river discharge introduces substantial complexity to the shelf circulation. Peak discharge from the Klamath River exceeds that of the Eel River discharge by 5,000 m$^3$s$^{-3}$. Nonetheless, due to a variety of factors including the easily erodible rock and steepness of the Eel drainage basin, the Eel River carries the largest sediment load of the California rivers (Sommerfeld and Nittrouer, 1999).

4.4 STATISTICS OF SHELF CIRCULATION

The G65 and S60 sites (over the 65 m and 60 m isobaths, respectively) on the northern California shelf were instrumented with tripod-mounted RDI workhorse ADCPs during the major January 1997 flood event (Ogston et al., 2000). The sites have an alongshore separation of about 18 km. Here the observations from those instruments are used to assess the performance of the 1-km and 3-km model simulations. All modeled and observed values are averaged over an inertial period (18 hours) and recorded every six hours. Modeled and observed velocities are then
rotated into the principal axes of the depth-averaged current. This effectively isolates the alongshore (major axis) and across-shore (minor axis) directions when the coastline has a variable orientation.

Observations at the S60 site are described by Ogston et al. (2000). The observed time-mean vertical structure of the major axis (alongshore) current at S60 (Fig. 4.3) is equatorward during the 32-day period surrounding the flood event. The standard deviation of the observed major axis (alongshore) current varies from 24 cm s\(^{-1}\) at 8 m depth to 14 cm s\(^{-1}\) at 58 m depth. The 1-km simulation vertical structure of the time-mean alongshore velocity is virtually identical to the observed structure. By contrast, the 3-km solution, though accurately tracking the observed S60 time series over 100 days (Chapter 3), does not reproduce the time-mean observed equatorward flow at all depths during the 32-day period.

In general, vertical profiles of standard deviation of the model alongshore velocity at the S60 site match the observed shape. In terms of magnitude, both the 3-km and 1-km solutions underestimate the variability near the bottom by about 5 cm s\(^{-1}\) while the 1-km solution overestimates the variability near the surface by about 5 cm s\(^{-1}\).

The observed time-mean alongshore velocity at the G65 ADCP site is equatorward at most depths (Fig. 4.3). Observed standard deviation of velocity at 7 m depth is 32 cm s\(^{-1}\) while it is reduced to 18 cm s\(^{-1}\) at 60 m depth. The velocity from the 1-km model reproduces the magnitude of the time-mean better than the 3-km model. In addition the 1-km model comes closer to matching the magnitude of observed velocity standard deviation above about 45 m than the 3-km solution. Observed variability is larger at all depths at G65 compared with S60, and the difference between them is greater near the surface than it is at depth. The relative difference in the standard deviation between the two
FIGURE 4.4 Time series of modeled and observed major axis (alongshore) and minor axis (across-shore) velocity at three depths at the G65 site. The ADCP data are shown courtesy of C. Friedrichs.
FIGURE 4.5 Surface velocity vector mean and rms vector amplitude of the 1-km resolution model over the 40-day simulation. Means are denoted with arrows, while rms amplitude is color contoured. Every sixth vector alongshore is shown, while every second vector in the across-shore direction is shown. The full north–south extent of the model domain is represented; however, the domain is truncated to the west to highlight the shelf region.
FIGURE 4.6 As in Fig. 4.5, but for depth-averaged velocity.
sites is qualitatively represented by both models. However, both models produce smaller standard deviations near the bottom compared with observations. The large spatial variability in mean and fluctuating velocity between G65 and S60 could contribute to discrepancies between the model and observations.

Observed and 1-km modeled minor axis (across-shore) time-mean and standard deviations at S60 and G65 (Fig. 4.3) are small in comparison to the corresponding major axis quantities. The 3-km model results are qualitatively similar to those from the 1-km model and are not plotted. The 1-km solution time-mean across-shore velocities are small in magnitude, but have differences in vertical structure compared with the observations. In particular, the observed time-mean velocity is onshore at S60 above 52 m depth while the 1-km solution presents a near-zero, depth-independent mean. The observed standard deviation is relatively barotropic in nature, with near-surface standard deviations exceeding those in the rest of the water column by less than 2 cm s\(^{-1}\). Modeled across-shore velocity fluctuations have stronger shear in the vertical than the observed fluctuations.

At G65 the observed across-shore time-mean current is directed off-shore at all depths. Magnitudes of the mean are larger than at the S60 site. Observed standard deviations are twice as large at G65 as they are at S60. The vertical structure of observed fluctuations is barotropic. As at S60, modeled across-shore current mean values have similar magnitudes, but differ in vertical structure, while the standard deviations are somewhat smaller than the observed values and have greater vertical shear. The pronounced differences in observed and simulated across-shelf velocity at the two sites separated by less than 20 km are presumably related to the relatively short alongshore correlation scales of across-shore velocity that are in contrast to the longer correlation scales of the alongshore velocity. The correlation between the S60 and G65 sites of observed
minor (major) axis velocity at 30 m depth is -0.22 (0.86), while the 1-km model
minor (major) axis velocity correlation between the S60 and G65 sites at 30 m
depth is -0.28 (0.83).

The time series of modeled and observed alongshore velocity at G65 (Fig. 4.4)
demonstrate that the 1-km solution reproduces the observed equatorward flow
associated with the arrival of equatorward winds better than the 3-km solution.
This feature is also found in the S60 time series (not shown) and likely accounts for
the difference in mean flow profiles between the 1 and 3-km solutions. Both 1-km
and 3-km alongshore velocities are highly correlated with the observed velocity.
Correlation coefficients are between 0.80 and 0.85 for both the 1-km and 3-km
models at all three depths. Minor axis (across-shore) correlations of model and
observed time series are low and similar for the 1-km and 3-km solutions. An
exception is at mid-depth (30 m) where the 1-km solution attains a correlation
coefficient of 0.56 at S60.

The horizontal spatial structure of the 1-km model time-mean and fluctuating
currents is exhibited in maps. Surface current variability is greatest over the
northern Eel margin (from the Eel River to Trinidad Head) and north of Point St.
George (Fig. 4.5). Over the 40-day simulation period the magnitude of fluctua-
tions in those regions is typically 40-50 cm s\(^{-1}\). Time-mean surface velocities are
poleward and have magnitudes of about 20 cm s\(^{-1}\) over the northern Eel margin
and 50 cm s\(^{-1}\) north of Point St. George.

The depth-averaged time-mean current is equatorward over the northern Eel
margin (Fig. 4.6). This is in agreement with the depth structure at the mea-
surement locations S60 and G65 (Fig. 4.3) and extends that result to illustrate
the broad horizontal spatial swath of mean equatorward depth-averaged flow.
Low-frequency depth-averaged current fluctuations are strong (approximately
20 cm s\(^{-1}\) over the shelf precisely along the stretch of coast, the northern Eel margin, that is the locus of STRATAFORM observations. Thus, STRATAFORM samples a highly energetic flow field relative to the surrounding ocean. Chapter 3 revealed this high variability north of Cape Mendocino to be unrelated to river run-off, though freshwater input increased the magnitude of fluctuations in the surface waters. The localization of maximum variability in discrete regions suggests the influence of coastline morphology, specifically the promontories of Trinidad Head and Point St. George, in enhancing circulation variability in the surface and depth-averaged maps.

4.5 TIME EVOLUTION OF EEL RIVER PLUME

Salinity measurements at the G60 site (over the 60 m isobath) during the 40-day simulation period are compared with model quantities in Fig. 4.7. Salinity measurements from the Seacat sensor are described by Geyer et al. (2000). As in the previous section, the modeled and observed time series consist of inertial averages centered at 6-hour intervals. The most prominent feature of the near-surface salinity time series is freshening as the Eel River plume sweeps across the monitoring site. This happens in the models just after the winds rotate to an upwelling-favorable direction. The modeled low-salinity pulse precedes the observed pulse by about a day and a half. The modeled (1 km) river plume expands offshore after January 2 as Eel River discharge peaks (Fig. 4.8). With the winds providing reduced forcing relative to earlier times, the plume moves offshore on January 2–3 mostly through its own dynamics. Geyer et al. (2000) point out that the axis of the Eel River mouth is rotated about 20° from the alongshore direction. They suggest that this oblique orientation causes the momentum from
the Eel River flood to be injected predominantly alongshore (northward) rather than across-shore. In contrast, the model river mouth enters the shelf in the across-shore direction. This may result in a wider plume and account for the earlier appearance of the low-salinity plume signature at G60 in the model. By January 4 the 1-km model plume is advected to the south of the G60 site under the influence of equatorward wind forcing (Fig. 4.8). The 1-km model salinity anomaly is about 13 psu and is closer to the observed plume in magnitude and duration than the 3-km model.

Temperature at 1 m depth was also measured at the G60 site (Geyer et al., 2000). Modeled fluctuations in near-surface temperature do not match the observed variability. Absolute errors, however, are small. The differences in standard deviation are 0.17° for the 3-km model and -0.03° for the 1-km model. Modeled mean temperature is slightly warmer than the observed mean. The differences in mean values are 0.47° for the 3-km model and 0.29° for the 1-km model.

The time series discussed above supply information at a single location. With an understanding in hand of the model strengths and limitations in reproducing time series observations during the flood, we next examine the model evolution of the across-shelf structure of the plume during the flood as represented by the 1-km model. All variables shown in subsequent sections are averaged over an inertial period.

Potential density (\(\sigma_\theta\)) and north-south velocity (\(u\)) on east-west sections extending through G60/G65 (on the G-line), near the mouth of the Eel River (Fig. 4.9), and through S60 (on the S-line), 18 km to the north (Fig. 4.10), are highly variable during the course of the Eel River flood. During strong poleward winds and in the initial flood stages (December 30) the shelf waters out
to about 10 km from the coast are relatively unstratified except for the surface plume. The developing plume is deeper and more bottom-attached on the S-line relative to the G-line. On both sections a poleward coastal jet over the shelf is evident. However on the S-line the jet is more tightly confined to the shelf than it is on the G-line. Velocity is poleward over most of the shelf at both sites; yet, there is equatorward flow near the bottom on the S-line inner shelf. During declining poleward winds and near-maximum river discharge (January 1) the plume thickens vertically on both sections. The poleward jet at both sites moves offshore. During equatorward winds and post-peak discharge (January 3), the plume spreads offshore and thins vertically at both sections. Over the outer shelf and shelfbreak the density attempts to restratify at both sections in response to the upwelling-favorable winds. The modeled Eel River plume response to the winds on the S-line and G-line mimics the across-shelf narrowing and vertical thickening (downwelling-favorable winds) replaced by across-shelf spreading and vertical thinning (upwelling-favorable winds) of the plume documented by Geyer et al. (2000) from helicopter-lowered CTD surveys. The flow reversal on the S-line gains strength over time such that by January 3 the flow on the inner shelf is predominantly equatorward. Equatorward flow of smaller magnitude begins to occupy the shelf region on the G-line as well.

An alongshore section of potential density through the plume on January 1 at an offshore distance of 1.5 km from the coast is shown in Fig. 4.11. Most of the low-density plume water is confined to the upper 5 m. Very light water ($\sigma_\theta < 18 \text{ kg m}^{-3}$) extends about 10 km north of the Eel River mouth in the surface waters. Light water stretching to about 10 km north of the mouth is qualitatively consistent with data from hydrographic surveys by Geyer et al. (2000). The plume deepens as it flows north, leading to the bottom-attachment seen on the
S-line (Fig. 4.10). The thickening of the plume to the north during downwelling-favorable winds was observed by Geyer et al. (2000) during helicopter CTD surveys.

To help clarify the role of vertical turbulent mixing in the physical response of the shelf and plume waters to the applied time-dependent wind forcing, sections along the S-line of gradient Richardson number,

\[ Ri = \frac{-g \partial \sigma_e}{\rho_0} \left[ \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 \right]^{-1}, \tag{4.1} \]

twice the turbulent kinetic energy \( q^2 \) from the Mellor and Yamada (1982) level 2.5 turbulence closure scheme, and vertical kinematic diffusivity \( K_H \) are displayed in Fig. 4.12. On December 30 enhanced mixing over the outer shelf is evident in \( Ri \), \( q^2 \), and \( K_H \). Fields of vertical kinematic viscosity \( K_M \) are similar to those of \( K_H \). The near vertical contours of \( q^2 \) and \( K_H \) suggest the position of an outer shelf downwelling front that has developed under persistent strong poleward winds during the preceding days (Allen and Newberger, 1996). Over the inner shelf there is some slightly increased mixing in the upper 30 m. On January 1 as the Eel River discharge rises and the winds slacken, increased \( Ri \) values on the outer shelf indicate a stabilized water column. Fields of \( q^2 \) have increased magnitudes presumably as a result of increased vertical shear near the surface in the alongshore current. The large values of \( q^2 \) are not accompanied, however, by large values of \( K_H \). The relatively small values of \( K_H \) are caused by the effects of stable stratification in the plume through the structure functions in the Mellor and Yamada (1982) formulation and are suggestive of an absence of significant vertical mixing. An evaluation, discussed below, of the balance of terms in the salinity equation shows, however, that vertical diffusion plays a significant role in determining the structure of the salinity field under the plume as might be
anticipated from the large vertical gradients of salinity and the significant wind stress during this time period.

In order to explore the mechanism for plume deepening on the S-line seen in Fig. 4.10, we partition the salinity equation into the tendency \( \frac{\partial S}{\partial t} \), advection (including both horizontal and vertical advection), and diffusion (including both vertical and horizontal diffusion) terms and plot these on December 30 and on January 1 on an east-west section along the S-line (Fig. 4.13). Dominant features of the term balance on both days are described here. Relatively large positive values of the advection term near the surface within 15 km of the coast represent the northward advection of low-salinity plume water by the alongshore current. This is primarily balanced by vertical diffusion, which mixes higher salinity water from below into the near-surface region as indicated by negative values of the diffusion term. Vertical diffusion also mixes low-salinity near-surface water from the plume to depths of 20-30 m as shown by the positive values of the diffusion term. The latter process is primarily balanced by negative advection. Assuming that the dominant contribution to advection is from the alongshore velocity, the change in sign in the advection term near 10 m depth is consistent with the alongshore structure of the potential density in Fig. 4.11 which shows increasing \( \sigma_\theta \) near the surface and decreasing \( \sigma_\theta \) below 10 m as the distance increases northward from the river. The relatively small imbalance between advection and diffusion results in a negative tendency term which is largest near the offshore edge of the plume and which represents a net freshening of the near-surface water as the plume increases in strength. An examination of the term balances on an alongshore section 5.5 km from the coast indicates that the processes described above, involving major contributions from both advection and vertical diffusion,
occur robustly at alongshore locations between the G-line and S-line from 30 December through 1 January.

By the end of the storm, significant structure has developed in the shelf flow field. Walsh and Nittrouer (1999) identified an eddy in Advanced Very High Resolution Radiometer (AVHRR) satellite imagery on January 6, 1997. They interpret the image of the visible band (channel 1) minus the near infrared band (channel 2) as representative of turbidity. We have found that the eddy is also evident in the far infrared (channel 4) temperature-corrected (Pathfinder SST algorithm) NOAA-14 satellite data (Fig. 4.14). The Eel River water is up to several degrees cooler than the ambient shelf water, which provides a thermal signature as the river plume undergoes an anticyclonic wrapping just north of Cape Mendocino. The eddy feature was identified in Chapter 3 and found to be a robust response to storm passage, even in the absence of river run-off.

The impact of the shelf circulation on the Eel River plume in the 1-km model simulation is apparent in Fig. 4.8. During strong poleward winds (January 1) the plume (as delineated by surface salinity) hugs the coast and courses northward with surface velocities on the order of 1 m s\(^{-1}\). As the winds reverse direction through January 2 and January 3 the plume spreads offshore. During the period from January 4 to January 5, the low-salinity waters from the Eel River are wound clockwise into the developing eddy and move farther offshore. This eddy is approximately 50 km in diameter, in agreement with the size of the eddy imaged by satellite.

The spatial complexity of the bottom flow is highlighted in Fig. 4.15. Maps of bottom velocity during the storm show weak equatorward flow on January 1 over the inner shelf adjacent to S60. This flow gains strength by January 2 and spreads to the surface waters two days later (Fig. 4.8). The upward expansion with time
of the equatorward flow was identified in the S-line sections (Fig. 4.10). Here, the reversal in bottom currents is found to be localized between the region north of the K63 site and south of Trinidad Head. The establishment of the pressure gradients that are important in forcing the equatorward flow reversal may be seen in the plots of sea surface elevation in Fig. 4.16. High sea surface elevation south of Trinidad Head is a prevalent feature from January 1 to January 3. These high sea levels diminish as the eddy takes shape and migrates offshore on January 4 and 5. A comparable 1-km resolution model simulation was conducted with no river run-off to assess the contribution of river run-off to the flow reversal. In that simulation, explored in more detail below, elevated values of sea level still occur south of Trinidad Head, but the magnitudes are reduced by about 5 cm s\(^{-1}\). Without river run-off, a weakened eddy begins to form about a day later than the river-enhanced eddy.

Sections of potential density and north-south velocity along the S-line from the “no rivers” simulation (Fig. 4.17) are markedly different from the simulation with rivers (Fig. 4.10). Without rivers the inner shelf waters are well mixed vertically during the storm event. On January 1 the coastal jet is weaker by about 50 cm s\(^{-1}\) and positioned farther offshore than the buoyancy-augmented jet. There is a hint of developing equatorward flow on the shelf. However, it lags the strong equatorward flow on January 3 seen in the simulation with rivers (Fig. 4.10).

A section along the S-line of the north-south pressure gradient with and without rivers (Fig. 4.18) shows increased negative pressure gradient in the plume near the surface, and increased positive gradients (southward pressure gradient force) inshore compared with the simulation without rivers. The positive pressure gradients on the inner shelf are weak and barotropic without rivers but become
stronger and more baroclinic when buoyancy from the river plume is included. The strong equatorward flow on the shelf on the S-line that developed between December 30 and January 1 can be ascribed to pressure gradients set up by the plume.

Several centimeters of fine-grained sediment accumulated on the 60 m isobath between the sites identified as S60 and G60 in Fig. 4.15 (Wheatcroft and Borgeld, 2000) as a result of the flood. Ogston et al. (2000) analyzed sediment flux in the bottom boundary layer at the G65, K63 and S60 sites and found convergent flux between S60 and K63. This convergence is consistent with the velocity vectors of Fig. 4.15. Data from S4 electromagnetic current meters in the bottom boundary layer along the 60 m isobath during the flood event (Ogston et al., 2000) are compared with modeled velocities at the same locations in Table 4.1. Statistics are computed for a 12-day period surrounding the flood event. In both the modeled and observed data, the K63 site is the locus of largest poleward mean alongshore bottom current as well as largest standard deviation. At G65, the poleward mean flow is weaker and less variable in both modeled and observed time series. At S60, equatorward mean bottom flow leads to convergent bottom currents between S60 and K63 along the 60 m isobath in the model and in the observations.

From helicopter-sampled suspended sediment concentrations in the plume, Geyer et al. (2000) establish that between January 2 and January 3 sediment was lost from the plume. In terms of the model results described here, this information implies that sediment is sinking out of the poleward-flowing plume on January 2 into the equatorward flow near the bottom. The sheared flow that the sediment experiences could be important in moving sediment southward of where it might have accumulated if the water below the plume were quiescent.
FIGURE 4.7 Observed and modeled near-surface salinity at G60. Salinity measurements are shown courtesy of W.R. Geyer.

TABLE 4.1 Modeled and observed major axis (alongshore) velocity (cm s\(^{-1}\)) time series statistics at STRATAFORM G65, K63, S60 current meter locations approximately 1 m above the seabed (December 25, 1996 – January 6, 1997). (Observed data shown courtesy of C. Friedrichs and A. Ogston.)
FIGURE 4.8 Time evolution of surface salinity and velocity on the Eel margin. Every third vector is shown in the alongshore direction. Every vector is shown across-shore. STRATAFORM observation sites are indicated with crosses. NOGAPS wind stress at the location of Buoy #30 is shown for reference. The velocity scale changes between the upper and lower figures.
FIGURE 4.9 East-west sections of potential density ($\sigma_\theta$) and north-south velocity ($v$) through the STRATAFORM G-line (sites G65 and G60). Units are kg m$^{-3}$ for potential density and cm s$^{-1}$ for velocity. Dotted lines represent southward flow.
FIGURE 4.10 As Fig. 4.9, but for east-west sections through the S-line (S60 site).
FIGURE 4.11 Alongshore section of potential density ($\sigma_\theta$) on January 1, 1997 at an offshore distance of 1.5 km from the coast. Units are kg m$^{-3}$. The section is 25 km long and extends north from the mouth of the Eel River. The locations of the STRATAFORM G-line and S-line are indicated.
FIGURE 4.12 East-west sections of gradient Richardson number $Ri$, twice the turbulent kinetic energy $q^2$, and vertical kinematic diffusivity $K_H$ along the S-line. Fields are in units of $m^2 s^{-2}$ multiplied by $10^2$ for $q^2$, and $m^2 s^{-1}$ multiplied by 10 for $K_H$. Contours of $Ri$ are 1, 0.5, 0.25 and 0.
FIGURE 4.13 Dominant terms in the salinity equation on an east-west section through the S-line. The terms are multiplied by $10^5$ and units are psu s$^{-1}$. Negative values are shown with a dotted line.
FIGURE 4.14 AVHRR satellite (NOAA-14) image collected by the GLOBEC Northeast Pacific program. Processing of the image was carried out by Ocean Imaging, Inc. The image shows the eddy just north of Cape Mendocino.
FIGURE 4.15 Time evolution of velocity (cm s$^{-1}$) on the bottom sigma level during the storm. Every third vector is shown alongshore. Every across-shore vector is shown. The 50 and 100 m isobaths are drawn for reference.
FIGURE 4.16 Time evolution of sea surface elevation on the Eel margin. Units are cm. Details are as in Fig. 4.8. The 50 and 100 m isobaths are shown in grey.
FIGURE 4.17 Same as Fig. 4.10, but for the simulation without rivers.
FIGURE 4.18 East-west sections of north-south pressure gradient, $\frac{1}{\rho_0} \frac{\partial \rho}{\partial y}$, through the S-line. Sections for the simulation with and without rivers are shown. Pressure gradients are multiplied by $10^7$ and units are m s$^{-2}$. 
With regard to classifying the dynamics of the Eel River plume during the flood event, the highly time-dependent behavior resulting from the extreme variability in the wind stress forcing and the river discharge between December 27 and January 6 (Fig. 4.2) makes meaningful parameter estimation and utilization uncertain. Nevertheless, a calculation and comparison of some relevant instantaneous scales should help provide context. Consequently, we proceed and choose January 2 when the wind stress is relatively weak and the discharge is high (Fig. 4.8) for some parameter estimates. The Rossby radius of deformation $R_D$ for the first baroclinic mode has been calculated at a number of locations in the plume by using the local stratification and water depth from the model, assuming a flat bottom, and solving the vertical eigenvalue problem. We find $R_D = R_{Di} \approx 10.5$ km near the river mouth and $R_D = R_{Dp} \approx 12 - 14$ km in the plume at distances of 4-10 km offshore and 10-15 km north of the river mouth. The corresponding internal wave velocities $c = fR_D$ are $c = c_i \approx 1$ ms$^{-1}$ near the mouth and $c = c_p \approx 1.15 - 1.33$ ms$^{-1}$ in the plume. From the model results we find that the offshore surface velocity near the river mouth is approximately $u_i \approx 1.2$ ms$^{-1}$ while the alongshore surface velocities in the plume are typically $v_p \approx 1$ ms$^{-1}$. The across-shelf scale $L_0$ of the plume, estimated from the surface salinity field in Fig. 4.8, is $L_0 \approx 12$ km. With these values, we estimate for January 2 the inlet Froude number $F_i = u_i/c_i \approx 1.2$, corresponding to weakly supercritical conditions in the inlet, and a plume Froude number $F_p = v_p/c_p \approx 0.75 - 0.87$, corresponding to subcritical flow in the plume. An inertial scale $L_I = u_i/f \approx 13$ km is close to $L_0, R_{Di}$ and $R_{Dp}$. The plume Kelvin number (Garvine, 1995), $K = L_0/R_{Di} \approx 1.2$. In addition, the Rossby numbers $\epsilon_i = u_i/fL_0 \approx 1.05$ and $\epsilon_p = v_p/fL_0 \approx 1.0$. Given the uncertainties inherent in these scale estimates, we feel that the most reasonable conclusion here is that for
January 2, the values of parameters $K$, $L_1/L_0$, $F_i$, $F_p$, $\epsilon_i$, and $\epsilon_p$ are all of order one. This implies that stratification, rotation, and nonlinear advective effects are all important contributing factors in the dynamical balances of the instantaneous Eel River plume on January 2.

4.6 DISCUSSION AND SUMMARY

This study has described results from a 1-km resolution nested model simulation of the northern California margin. The simulation focused on the major Eel River flood event which occurred as a severe storm passed over and generated winds that veered from poleward (downwelling-favorable) to equatorward (upwelling-favorable). High-resolution modeling enables the mapping of statistical properties of flow variables. Such maps of mean and rms fluctuations of currents are valuable tools for characterizing the dynamic coastal ocean. The shelf flow on the northern Eel margin during the 40-day model simulation consists of strong mean poleward surface velocities with rms fluctuations of 50 cm s$^{-1}$. Modeled depth-averaged time-mean flow is equatorward with rms fluctuations of 20 cm s$^{-1}$. Though possessing oppositely directed time-mean currents, both surface and depth-averaged flow display enhanced intensity over the northern Eel River margin.

Observed time-mean major axis (alongshore) velocity at S60 is equatorward at all depths, while the observed standard deviation of the alongshore velocity is on average 20 cm s$^{-1}$ in the water column. The 1-km simulation reproduces the vertical structure of the mean alongshore flow over the shelf at S60 better than the 3-km simulation. In agreement with the observations, fluctuations in alongshore and across-shore 1-km model velocity are larger at the G65 site compared to
the S60 site. Major axis (alongshore) 1-km and 3-km model currents are highly correlated (0.80 ≤ r ≤ 0.85) with observations at G65. Model simulations of minor axis (across-shore) velocities are not as successful, presumably due in part to the short correlation scales of across-shore velocity.

Across-shelf sections during the flood show a near-surface plume advancing onto an extremely well-mixed shelf. Deepening of the plume to the north between the G-line and S-line is a prominent feature of the alongshore plume structure during downwelling-favorable winds. This feature appears to be associated primarily with vertical mixing that is presumably intensified by the large wind stress. Properties of the modeled river plume agree with observations taken during downwelling conditions (Geyer et al., 2000). Under the influence of upwelling-favorable winds the plume thins and stretches offshore. The modeled plume sweeps across the G60 site several days earlier than the observed plume as evidenced by a comparison of salinity time series. This discrepancy may be due to the model river plume being wider since the model inlet geometry constrains the resultant momentum from the river to be input to the shelf in the across-shore direction. The axis of the Eel River mouth, however, is rotated about 20° from the alongshore direction so the river would input momentum northward.

In satellite imagery and model simulations, an anticyclonic eddy forms over the outer shelf as the storm passes and winds shift equatorward. Based on the 3-km model simulation, Chapter 3 demonstrated the eddy to be a generic feature of the relaxation response from strong poleward wind forcing during storms, whose circulation is enhanced by river run-off. The low-salinity Eel River plume water becomes entwined in the developing eddy. The onshore limb of the eddy is linked to equatorward flow over the inner shelf. Bottom velocities over the inner shelf turn equatorward several days before the surface velocities in the 1-km simulation.
The equatorward flow over the shelf during the transition of winds from strongly poleward to weakly equatorward is attributed to a southward pressure gradient force that develops south of Trinidad Head during strong poleward winds. This pressure gradient force is both intensified and baroclinic in structure due to the presence of the Eel River plume.

The observed suspended sediment concentrations, sediment flux vectors and sediment accumulation on the 60 m isobath (Geyer et al., 2000; Ogston et al., 2000; Wheatcroft and Borgeld, 2000) may be indicative of sediment leaving the fast northward-flowing plume over the inner shelf, up to 60 km from the river mouth (Geyer et al., 2000), and sinking through a highly sheared water column with southward flow at depth. Convergent bottom currents in the model and observations steer flow to the region between S60 and K63. Hence sediment accumulation patterns may be linked to these low-frequency circulation patterns.

There are many documented instances of persistent circulation features amassing sediments in preferred locations (Dyer and Huntley, 1999). The eddy described here is a transient phenomenon, occurring in response to passing storms. The extent to which the low-frequency circulation could determine the fate of sediment may best be approached by a combined sediment transport/physical circulation model that is cognizant of the major role played by coastline irregularities, wind forcing, and river run-off in determining the shelf response, as demonstrated here.

4.7 ACKNOWLEDGMENTS

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4.8 REFERENCES


In Chapter 2, the performance of nesting techniques was evaluated in a simple model configuration. Here, a brief discussion of nesting in the context of the realistic simulation presented in Chapters 3 and 4 is undertaken.

Near the boundary between the NRL PWC (9 km) model and the 3-km first nested model there were some steep gradients in the mean fields of surface velocity and surface elevation (Figs. 5.1 and 5.2). The mean values and the standard deviations of the surface velocity and surface elevation from the 3-km first nested model and from the 1-km second nested model over a common time period are shown in Figs. 5.3 and 5.4. The gradients at the boundary between the 3-km and the 1-km grids are reduced considerably compared to those between the 9-km and 3-km grids. Mismatches between the 3-km and 1-km mean surface elevation at the boundary typically do not exceed 0.5 cm. Furthermore, eddies positioned in the mean at the interface between the grids were not severely distorted by the transition between grid resolutions - e.g., at the southern boundary next to the coast in Fig. 5.4. By contrast, distortion of mean eddies was evident in the transition between the 9-km and 3-km models, especially near the western (offshore) boundary. Mean surface velocity fields were also better matched at the boundary between the 3-km and 1-km models compared with the boundary between the 9-km and 3-km models. Noisy spatial oscillations at the nested grid boundary were most pronounced in the velocity component tangential to the boundary near regions of outflow from the inner grid. These oscillations were markedly reduced in the 3-km to 1-km nested simulations.

The superior performance at the boundary between the 3-km and 1-km model might be attributable to the compatibility in bathymetry between the 3-km and
1-km models. These bathymetry fields were constructed to be conservative, as defined in equation (2.1), by first generating a 1-km bathymetry over a large domain that was then averaged to create bathymetry on the 3-km resolution grid. Bathymetry on the 3-km model at the boundary with the 9-km model was blended linearly over five cells to match the 9-km bathmetry, conservatively interpolated, at the boundary. Interior to the boundary cell, the bathymetry on the 9-km and 3-km domains come from different datasets. Apart from these differences in bathymetry generation, the nesting procedure was the same for the two nests, as summarized in section 3.3.

As discussed in Chapter 2, it seems prudent to position the region of dynamical interest at a distance from the nested boundary, as enacted here. This serves to reduce the impact of boundary reflections on the interior dynamics. In addition, efforts to match bathymetry inside the boundary in order to preserve consistency appear to have been rewarded, as documented above.
FIGURE 5.1 NRL PWC (9 km) and first (3 km) nested model surface velocity mean and standard deviation for the 40-day period of the second nested model simulation. Means are shown with arrows while standard deviation is color contoured. The color scale common to both panels is shown in the right panel. Units are cm s\(^{-1}\).
FIGURE 5.2 NRL PWC (9 km) and first (3 km) nested model surface elevation mean and standard deviation for the 40-day period of the second nested model simulation. Means are contoured in black while standard deviation is color contoured. The color scale common to both panels is shown in the right panel. Units are cm.
FIGURE 5.3 First (3 km) and second (1 km) nested model surface velocity mean and standard deviation for the 40-day period of the second nested model simulation. Means are shown with arrows while standard deviation is color contoured. The color scale common to both panels is shown in the right panel. Units are cm s$^{-1}$. 
FIGURE 5.4 First (3 km) and second (1 km) nested model surface elevation mean and standard deviation for the 40-day period of the second nested model simulation. Means are contoured in black while standard deviation is color contoured. The color scale common to both panels is shown in the right panel. Units are cm.
6 SUMMARY

High-resolution modeling studies of shelf circulation on the U. S. West Coast are scarce. Those that exist address circulation features associated with upwelling conditions. Yet winter contains a rich array of forcing mechanisms whose impact on shelf circulation has not been explored in prior modeling studies. In particular, equatorward winds that typify the spring/summer season are replaced by veering winds in winter as storms sweep across the Pacific and pound the northern coastal regions of California. In addition, flooding of coastal rivers by precipitation derived from the severe storms creates a substantial input of momentum and buoyancy to the coastal ocean. Recently, observational analysis has begun to focus on the unique dynamics of wintertime circulation as a topic deserving of exploration (Lentz and Trowbridge, 2000). And ongoing STRATAFORM observations over the shelf off northern California are sampling the challenging wintertime environment created by the combined action of winds, rivers and complicated terrain (Nittrouer, 1999). These observations set the stage for a realistic high-resolution simulation of northern California winter coastal circulation by providing data against which model results can be validated.

This thesis undertakes a series of high-resolution studies of continental shelf processes. In order to account for forcing that may originate from outside a local region, the numerical finite-difference method of "nesting" is adopted. Nesting is a means of embedding a high-resolution domain inside a lower resolution domain. Techniques for carrying out the communication between grids of differing resolution are tested in Chapter 2 using an idealized model domain and forcing that produces circulation features that characterize continental shelf flow. These features include time-dependent upwelling fronts and surface and bottom
boundary layers. The nested grid has 1 km resolution while the surrounding grid has 3 km resolution. A radiation nesting method (Carpenter, 1982) performed best in a simulation having no variations in forcing alongshore, as well as in a simulation that had spatially variable wind forcing in the alongshore direction. Though superior in the idealized tests conducted in Chapter 2, radiation nesting proved less robust than "clamped" nesting in realistic simulations. The radiation approach required an unreasonably small time-step in preliminary simulations of the coastal domain off northern California. In addition, the clamped technique, though inferior to radiation nesting in the test studies, generated solutions that were comparable to the radiation technique within 10 km of the downstream nested grid boundary under alongshore varying forcing conditions. Thus, clamped nesting was used for the realistic simulations described in Chapters 3 and 4. Incorporating the lessons from the idealized simulation, the downstream (northern) boundary of the realistic nested grid domains was extended farther north in relation to 40.6°N, the approximate locus of STRATAFORM observational data (Fig. 3.1).

Chapter 3 describes the first nested model (3 km resolution) simulation of winter 1996–1997 shelf circulation in the vicinity of Cape Mendocino. This model grid is embedded in the Naval Research Laboratory's Pacific West Coast model (NRL PWC). The NRL PWC is an approximately 9 km resolution regional model of the North Pacific Ocean. Model forcing consisted of realistic winds, river run-off from several coastal rivers, and high-resolution coastline and bathymetry. The latter traces a highly convoluted path in the region around Cape Mendocino and is an additional forcing mechanism on this stretch of the northern California coast. The 100-day simulation was also conducted without the river forcing in order to isolate wind-forced dynamics from buoyancy-driven dynamics. The sim-
ulations were used to statistically describe winter circulation. Mean flow north of Cape Mendocino was poleward in the upper 40 m over the shelf (with and without rivers). In agreement with historical NCCCS observations from winter 1988–1989, near surface mean flow south of Cape Mendocino was equatorward. Other similarities include stronger alongshore polarization of currents south of Cape Mendocino during NCCCS and in the 1996–1997 model simulations. This difference in polarization north and south of Cape Mendocino is consistent with a broad shelf north of Cape Mendocino compared to the narrow shelf south of the cape. Probably as a result of the uncharacteristic dearth of winter storms in winter 1988–1989, circulation differences exist in mean velocity over the shelf north of Cape Mendocino.

Model surface velocity variability is larger north of Cape Mendocino compared to the region to the south. Variability is strongly enhanced in two distinct regions north of Cape Mendocino when rivers are included in the simulation. This variability is surface intensified (upper 10 m) and is associated with a strengthened poleward jet over the shelf when river run-off is included.

Nested model velocity time series at several depths show strong agreement with observed velocities on the 60 m isobath over the shelf (averaged through the water column: $r=0.78$) compared with the coarse resolution “outer model” (averaged through the water column: $r=0.5$). On the nested grid, this midshelf site responds to storm passage by a pattern of fluctuations in depth-integrated acceleration and bottom stress. An equatorward pressure gradient force aids in the current reversal that occurs as the trailing edge of the storm passes over the region. This current reversal over the shelf is associated with the inshore arm of an anticyclonic (clockwise) eddy that forms off the Eel River during the two major storms of winter 1996–1997. Though positioned adjacent to the river, the
eddy was also present in the simulation with river run-off turned off. This robust circulation feature is located over the continental slope and extends to depths of 500 m. By contrast, an eddy adjacent to the Klamath River only occurred when river run-off was included in the model. That feature was located over the continental shelf and had a shallow depth structure. In EOFs computed for model surface velocity the eddy adjacent to the Eel River appeared as a mode 2 feature. This mode accounted for 13.81% of the mean product and had an amplitude that increased in strength as winds relaxed and reversed following intense poleward storm-generated winds.

Chapter 4 detailed the results of a 1-km resolution simulation of the 80-year Eel River flood event of early January 1997. During a 40-day period containing the major flood, depth-averaged mean flow is equatorward over the Eel margin shelf area (in agreement with the midshelf mean vertical profile at STRATAFORM observational site S60) with intense rms magnitude of current fluctuations in that locale. Mean surface velocities are strongly poleward (20–50 cm s⁻¹) over the shelf. Fluctuations show enhanced intensity over the Eel River margin. Both surface and depth-averaged flow display strong alongshore variations over the shelf. Time series of modeled (1 and 3 km) major axis (alongshore) velocity over the midshelf G65 site show high correlation with observed velocity. During downwelling-favorable winds the plume thickens vertically as it flows northward. During upwelling-favorable winds the plume thins vertically and moves offshore. Characteristics of the plume structure agree with observations during the flood (Geyer et al., 2000). An anticyclonic eddy that forms over the shelf as the winds reverse is related to equatorward flow over the inner shelf that develops first at the bottom and subsequently expands through the full water column. The low salinity Eel River plume water becomes wound into the devel-
oping eddy. The occurrence of this low-frequency shelf circulation coincides with the time of highest sediment deposition and may play a role in moving sediments to their observed location.
BIBLIOGRAPHY


